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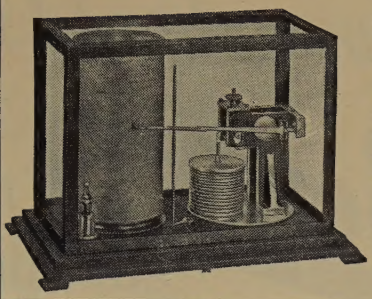

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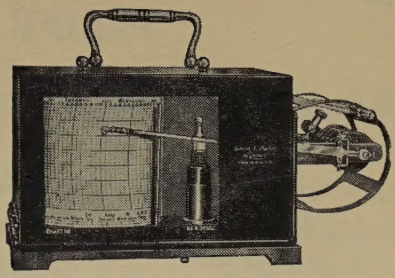

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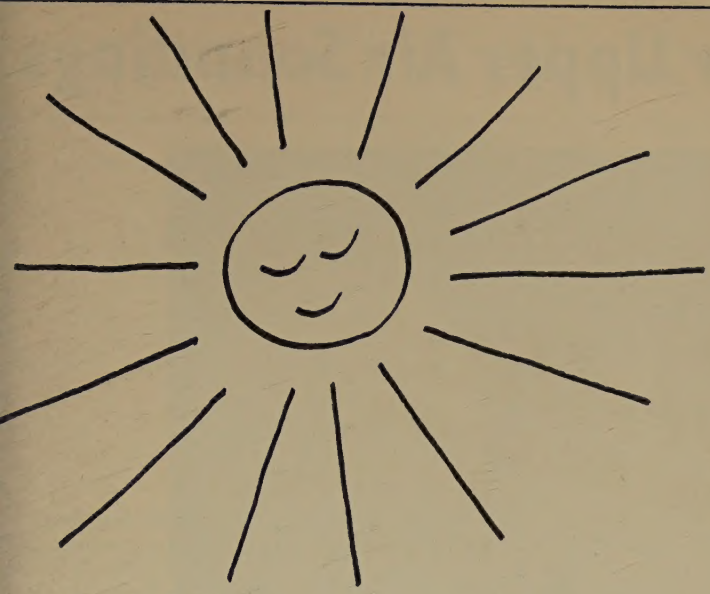

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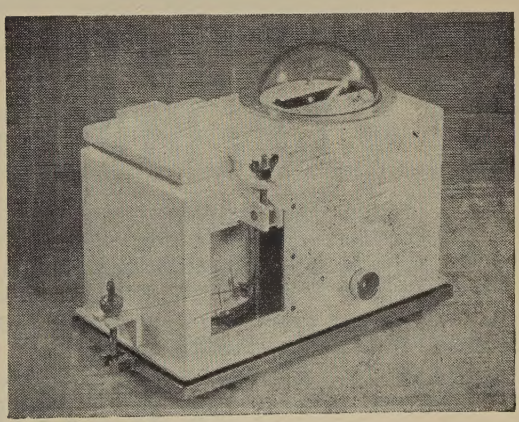

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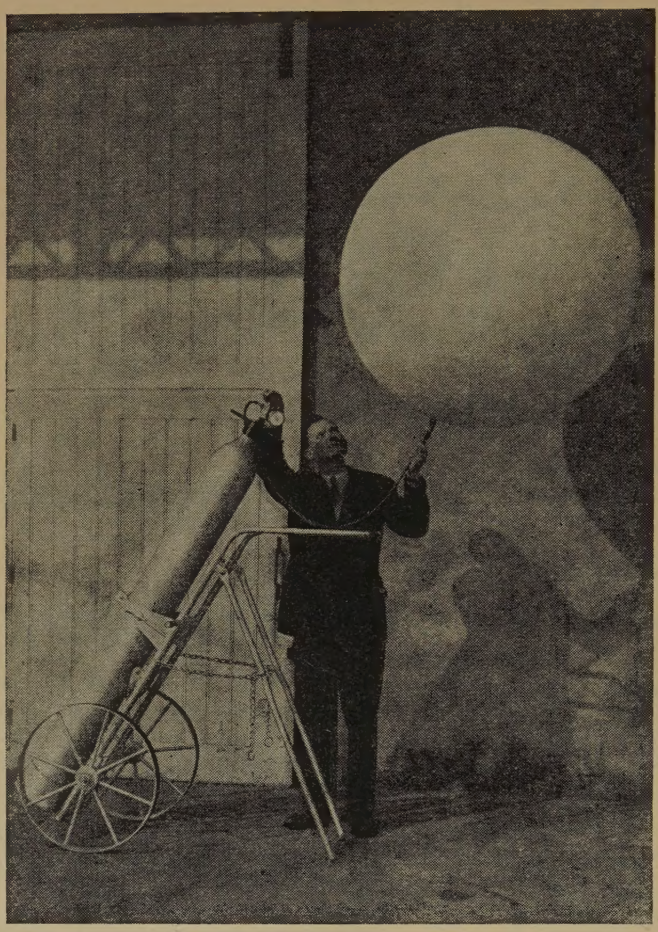

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The southern westerlies : a preliminary survey; main characteristics and apparent associations

By H. H. LAMB
Meteorological Office, Harrow

(Manuscript received 15 January 1958; in revised form 2 September 1958)

SUMMARY

Provisional figures can now be given for the ratio of momentum in the southern circumpolar westerlies to the momentum of the better known northern system. This ratio appears to be 1·5, or rather more, over the year as a whole and rises to over 4 in the northern summer-southern winter.

The preponderance of the southern system is such that it may be regarded as the ' flywheel ' of the general atmospheric circulation and there are indications that changes in the southern westerlies entail world-wide adjustments.

Preliminary survey at the 500 mb level shows the main trough south of 40°S centred about 100°-110°E at most seasons of the year and the principal ridge in the neighbourhood of the outer reaches of the Ross Sea. The mean troughs and ridges have notably smaller amplitude than in the northern hemisphere. Control of trough and ridge positions appears to be thermal rather than orographic, associated with the position of the broadest part of the Antarctic cooling surface. Little seasonal movement of the troughs and ridges in longitude is discerned and no very great change of circulation intensity, but there is an intensification of flow and increase of amplitude in the Indian Ocean trough from summer to late winter.

The pattern of glaciation, as well as available data on depression tracks and anticyclone distribution over the Southern Ocean, support the pattern of ridges and troughs at 500 mb here described and throw light upon the occurrence of blocking in various sectors of the southern hemisphere.

This study suggests that it may be the momentum and energy of the southern westerlies which are ultimately responsible for maintaining the meteorological equator on average north of the geographical equator, at least over the oceans.

In any climatic epoch when the southern westerlies were (i) farther south than now, and/or (ii) weaker than now, the meteorological equator should under this proposition have been nearer the true equator than is now the case. The work of other investigators has suggested that the equatorial low-pressure belt should also be wider at times when the circumpolar westerlies weaken in either hemisphere. The changes here postulated appear to fit in with what is known of conditions around, and since, the culminating phase of the ' Little Ice Age ' in the northern hemisphere about a century and a half ago.

1. INTRODUCTION : DATA AVAILABLE, MAGNITUDE AND INFLUENCE OF THE SOUTHERN CIRCUMPOLAR WHIRL

It has long been known that the southern hemisphere westerlies were stronger than their northern hemisphere counterparts. This knowledge is enshrined in the old name ' Roaring Forties,' scarcely applicable in the northern hemisphere. The great pressure difference between the sub-tropical anticyclones and the sub-Antarctic depressions early engaged the attention of meteorologists (Ferrel 1856, Murphy 1856, Hann 1915, Flohn 1944).

Data are still admittedly scanty. Owing to the vast extent of the southern oceans which occupy the zone of westerlies, this sparsity of direct observational material continues even during the International Geophysical Year, although material which has recently become available makes it possible for the first time to estimate the magnitude of the preponderance of the southern circulation.

It may be that the results of this preliminary survey will further efforts to improve the observation network, especially in the South Pacific, where introduction of weather-ship expeditions seems to afford the only hope of filling the biggest gap in our knowledge of the atmospheric circulation.

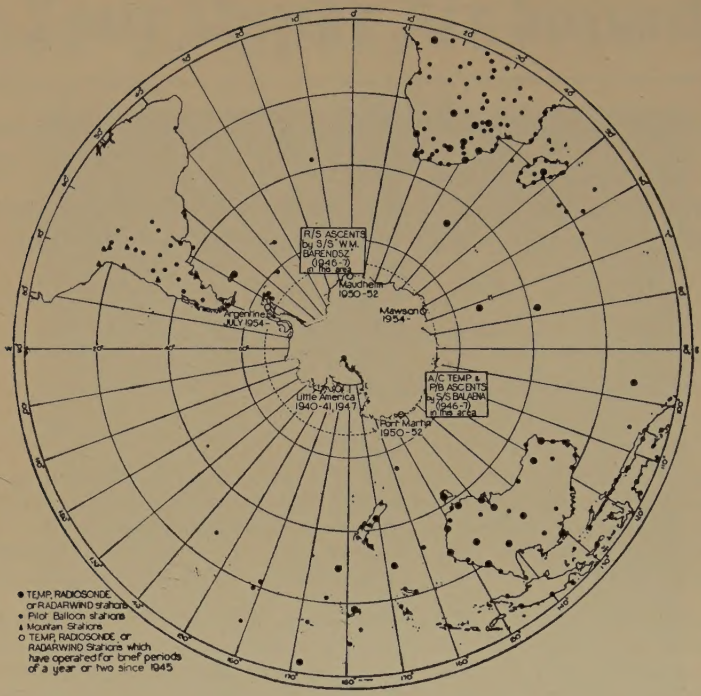
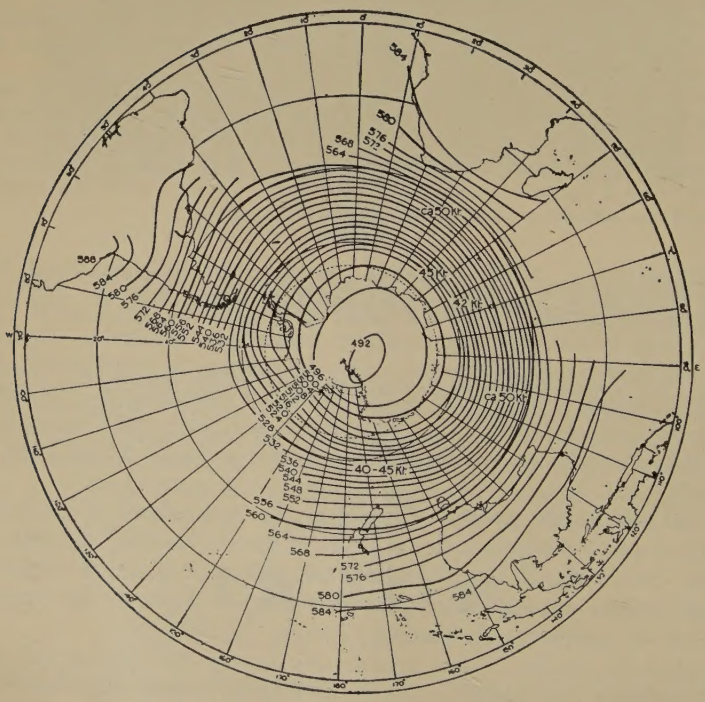


Figure 1. Network of aerological stations.



Monthly mean maps of the 500 mb surface over the southern hemisphere for each month from December 1951 to November 1954 have been published in *Notos* (Weather Bureau 1953-1955). The network of upper-air stations used (Schmitt 1953) is repeated here in Fig. 1, on which have been added indications of the few points on and near the Antarctic continent for which upper wind and temperature soundings have been available at some time since 1945. In the same article Schmitt has described the methods used in the daily hemispheric surface and upper-air analysis at Pretoria from which grid-point values are derived to give the monthly mean 500 mb pattern. The method included derivation of a basic 1,000 mb topography, and a map of 1,000-500 mb thickness, each chart plotted once a day with all data for that day which became available within the first six months after the event. The 500 mb topography was obtained by graphical addition, having regard also to the data plotted for that level and to the principle of continuity in time. The patterns of 1,000-500 mb thickness were completed with the aid of extrapolation of the eastward movement of identified features, subject to the control imposed by passing over test stations such as the islands of Marion, Heard and Macquarie with their regular radiosonde ascents. Use was also made of the discovery that at thoroughly oceanic stations the surface dew point in °F is nearly equal to the 1,000-500 mb thickness in dekametres with the 500 omitted. Corrections of 2 to 4 dekametres required in certain areas are systematic and the method appears sufficiently verified by internal controls, at least between South Africa and Australasia. For the eastern South Pacific, the broadest part of South America and the South Atlantic except between 40°-60°S, the working hypotheses used in constructing the daily charts had to be carried on without available checks within the region.

Whatever doubts might be entertained about the adequacy of the observation network for the individual daily or monthly maps, confidence has been increased by subsequent verifications available in respect of the daily surface maps (Lamb 1951, 1956); this must be attributed to some extent to the evident possibility of positioning correctly the large-scale features of the main steering currents in the upper atmosphere over the great Southern Ocean. This appears to be also the view of the South African Weather

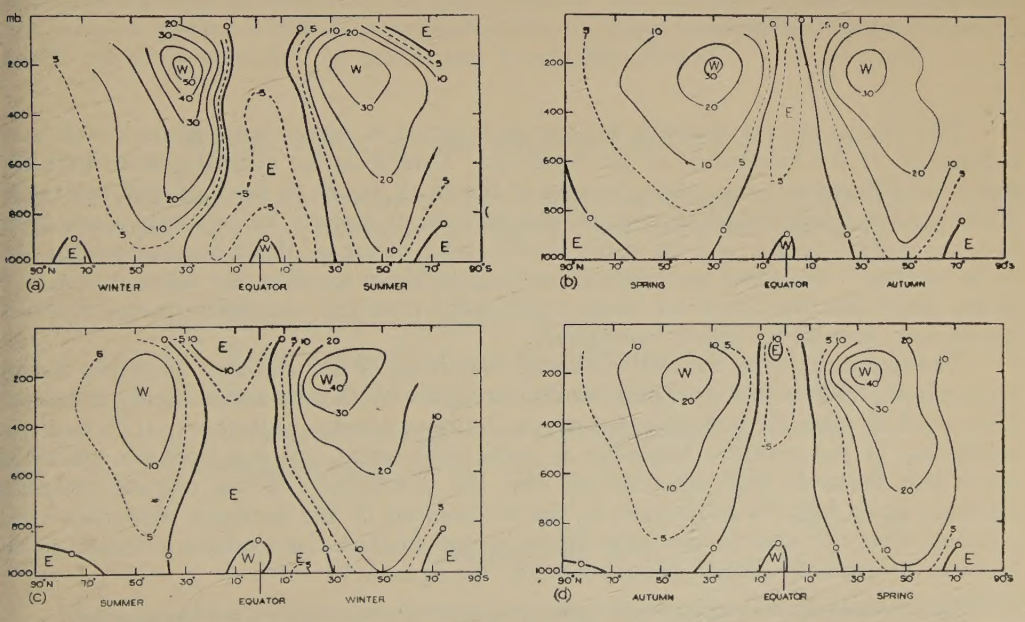


Figure 3. Cross-sections of the mean westerly component of velocity (used in momentum-ratio computations). (a) 30°-130°E in both hemispheres. (b), (c) and (d) All longitudes in northern hemisphere, 70°W-170°E in southern hemisphere (velocities in m sec^{-1}).

Bureau which has published the monthly mean charts in the belief, expressed by Schmitt (1956), that they hold possibilities for the study of at least the broad-scale features of the upper-air flow patterns, which should be exploited for research.

The three-year mean pattern yielded by the 36 monthly *Notos* values of 500 mb height and presented here in Fig. 2, can be used as a first approximation to the true picture. In Fig. 2 random errors have certainly been reduced to a small order, and the network hardly allows systematic errors in the large-scale features at least between 70°W and 170°-180°E. In any case we shall find it possible to confirm the main features from geographical evidence.

Figs. 3 (a) to (d) are approximate cross-sections from 90°N to 90°S of mean westerly velocity component in the four seasons. Fig. 3 (a), northern winter-southern summer, is for the sector 30°-130°E, which contains nearly the strongest upper westerlies in both hemispheres and at roughly constant latitude throughout the sector. In the other seasons longitudinal variations are not so great and the cross-sections have been made to relate to the whole northern hemisphere and the widest sector of the southern hemisphere for which enough data are available. The sections have been constructed from published maps of upper winds over the world (Brooks *et al.* 1950, Bannon 1954a, Jenkinson 1954), amplified by the southern hemisphere maps here given at Figs. 2, 6 and 7 as well as those for spring and autumn which are not shown. Mean cross-sections for the northern hemisphere 1941-1952 given by Moore (1956) have been extensively used.

A tentative integration of the westerly angular momentum in either hemisphere between latitudes 0° and 70° and from the 1,000 mb up to the 100 mb level has been made and the totals for each hemisphere suggest the following ratios (southern hemisphere/northern hemisphere) :

	Angular momentum of the westerlies Ratio	Net westerly angular momentum Ratio
S summer/N winter	0·7	0·6
S autumn/N spring	1·3	1·3
S winter/N summer	4·4	10·2
S spring/N autumn	1·6	1·9
Year	1·4	1·6

The figures for net westerly angular momentum in either hemisphere result from subtraction of the momentum of the easterlies. Extending the estimates to latitude 80° makes no difference to the ratios, because of the small extent of the high-latitude zones. Rather greater uncertainty is involved in the estimates of easterly and westerly momentum over the tropical oceans, and the entire Pacific sector of the southern hemisphere has had to be omitted; but these uncertainties are hardly likely to affect the ratios greatly. It seems likely that the ratios are more trustworthy than the necessarily rough totals of momentum derived for either hemisphere.

The figures point to the overwhelming magnitude of the southern westerlies during a considerable part of the year, particularly during the northern summer-southern winter. The very high ratios for June-July-August result from a number of factors : (i) the seasonal weakening of the northern westerlies is particularly great, (ii) the northern westerlies move into relatively high latitudes and (iii) the momentum of the tropical easterlies becomes quite large in proportion to the momentum of the northern westerlies – this situation does not arise in the same degree in the southern hemisphere summer, partly because the southern climatic zones are nearer the equator.*

* The following seasonal ratios of gross easterly to gross westerly angular momentum resulted from these calculations :

Northern hemisphere		Southern hemisphere	
Summer	0·58	Summer	0·18
Winter	0·09	Winter	0·02

The overweight of momentum in the southern westerlies is most marked in the lower troposphere between the surface and 500 mb; the result does not greatly depend on the (probably cautious) values here selected for 300 mb and higher levels, chiefly taken from Jenkinson (1954). Revised values for high levels over the Southern Ocean would probably raise the ratios slightly, so that the best values of the ratio for the southern summer/northern winter are likely to be in the region of 0·8 and for the year as a whole in the range 1·5 to 2.

The seasonal change is greater in the northern than in the southern hemisphere, the respective winter/summer ratios of angular momentum in the westerlies being 4·1 and 1·5. Greater seasonal change of latitude of the westerlies as well as of intensity in the northern hemisphere contributes to this result.

Angular momentum in the southern westerlies appears to be greater than in the northern westerlies over the longitude ranges defined in Figs. 3 (a) and 3 (c), summer for summer in the ratio 2·9 and winter for winter in the ratio 1·1. The overall average southern hemisphere/northern hemisphere ratio of westerly linear momentum is 2 to 2·5.

Schell (1939) has noted that G. T. Walker found more significant lag correlations between centres of action in the southern hemisphere and later conditions in the northern hemisphere than were found within the northern hemisphere itself. Lag correlations within the southern hemisphere were over three times more numerous. These discoveries may perhaps be regarded as expressions of the dominance of the southern westerlies.

It is reasonable to suppose that not only the zonal components but also the strongest meridional circulations should themselves show some relationship to the strongest upper westerlies. This is in accordance with experience in connexion with the winter jet stream of the northern hemisphere and the cold outbreaks over eastern Asia and north America and suggests a mechanism whereby the generally stronger circulation over the southern hemisphere can extend its influence to the northern hemisphere. Tucker (1957) found that encroachment of the southern-hemisphere meridional circulation north of the equator was a feature of most of the year in most longitudes. Correspondingly, impulses traceable in origin to the cold fronts of the Southern Ocean commonly cross the equator (Lamb 1957a, b) whereas similar impulses crossing the equator from the north are more restricted in season and longitudes affected.

2. Character of the southern circumpolar whirl

The prominent characteristics of the southern hemisphere westerlies (Fig. 2) about which positive statements can be made may be listed under the following heads :

(a) Amplitude of waves in the mean flow
(b) Mean trough and ridge positions
(c) Relationship of (b) to polar outbreaks
(d) Intensity and breadth of the main belt of westerlies
(e) Distinction of polar front and subtropical jet-streams
(f) Seasonal variation

(a) Amplitude of waves

Mean troughs and ridges can be identified (Fig. 2), but these are of modest amplitude by comparison with those of the northern hemisphere (Fig. 4) and are occasioned partly by successions of mobile features reaching their greatest amplitude near certain preferred longitudes.

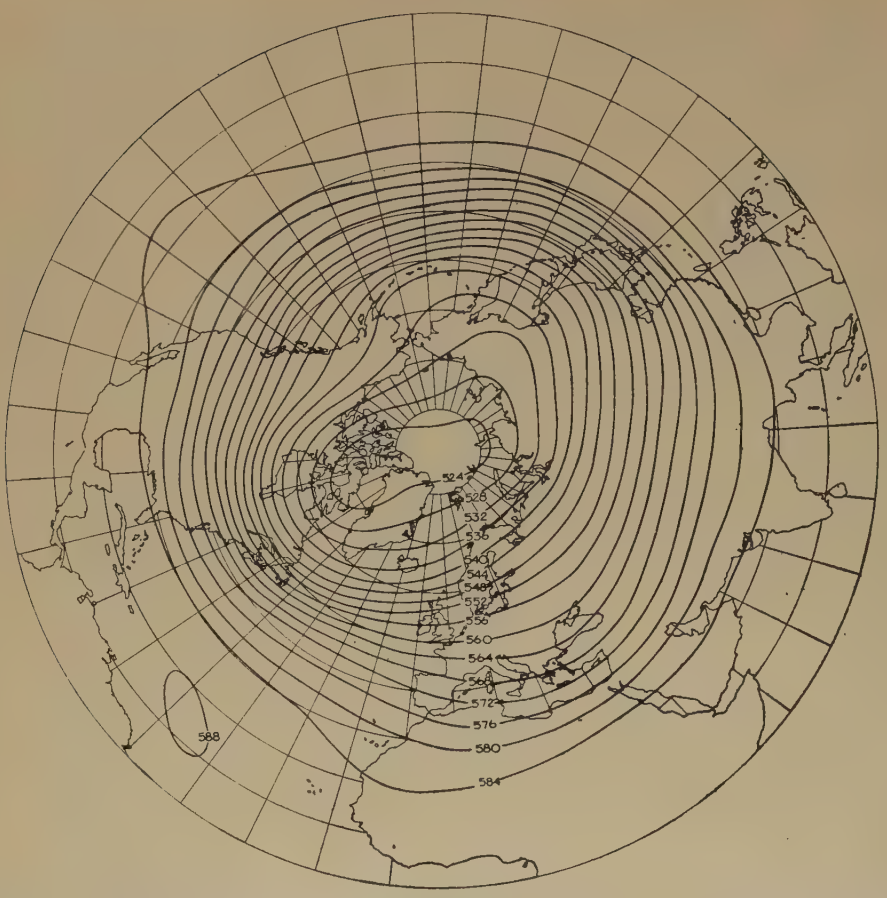


Figure 4. Mean height of 500 mb surface in dekametres for the months Jan. Apr. July and Oct. 1949-1953 (northern hemisphere).

(*b*) *Trough and ridge positions*

Identification of the positions of the mean troughs and ridges is most easily seen from profiles drawn along the latitude parallels (Fig. 5 (a)). Similar profiles for the individual seasons (Fig. 5 (b)) repeat most features of the overall average.

The most striking points are :

(i) In latitudes south of 45°S the trough downstream from the Andes is less pronounced than the broad trough in the Indian Ocean sector with its axis about 100°-110°E.* Between 20° and 45°S the trough in the western South Atlantic is possibly the biggest feature, in agreement with the deductions of J. A. Boffi (1949) from depression tracks and pilot-balloon data.

(ii) There is a ridge over the Andes to about 50°S, but from 50°-70°S (and possibly even north of 50°S) the ridge in the western Pacific-Australasian sector (150°E-150°W) is a much more noticeable feature.

(iii) There is a secondary ridge, unmistakable in 20°-50°S, between the longitude of South Africa and 40°-60°E.

* That the main trough is a broad feature occupying the Indian Ocean sector agrees with preliminary results reached by Schmitt (1952). According to Schmitt, and also to Heastie (1957), the axis of the trough in summer is rather in the western part of the Indian Ocean sector in longitudes 50°-70°E. The summer profiles at 40° and 60°S (Fig. 5 (b)) tend to support this position. In latitudes 30°-40°S the trough may split in summer into two troughs near 50° and 100°E. The all-year profiles (Fig. 5 (a)) for the lower latitudes 20° and 30°S also suggest a trough in the western part of the Indian Ocean sector (60°-90°E).

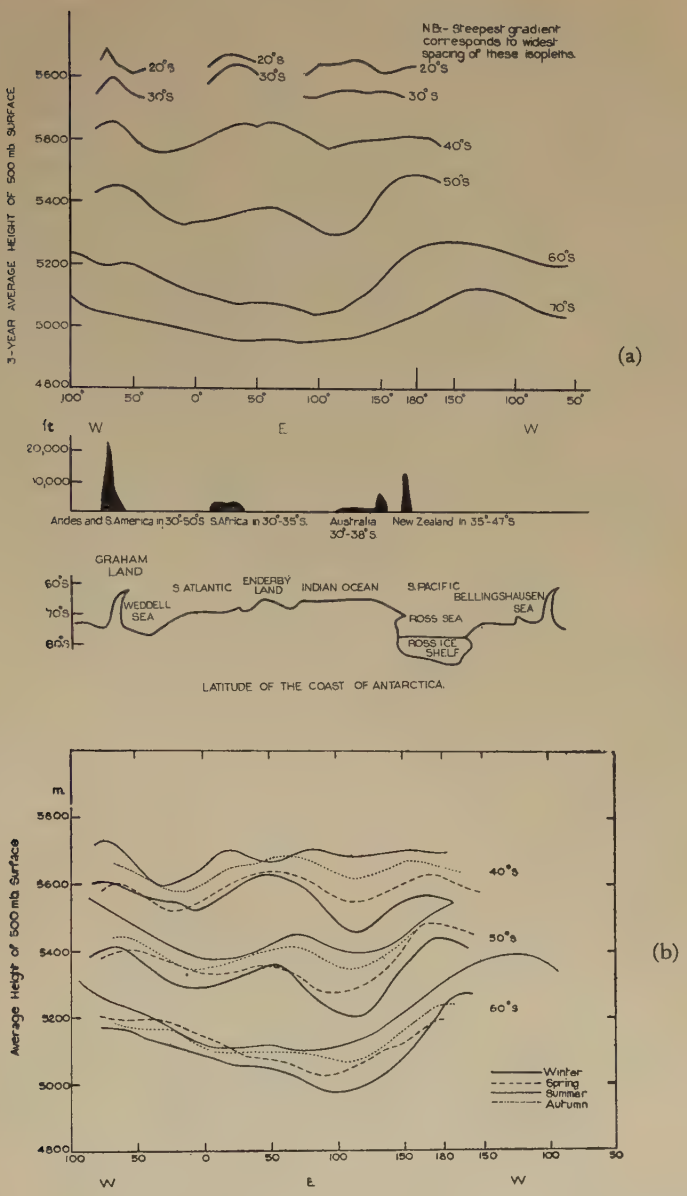


Figure 5. (a) 3-year mean profiles of the 500 mb surface along each 10° parallel of latitude (Dec. 1951-Nov. 1954); (b) Seasonal profiles of 500 mb height along 40°S, 50°S and 60°S (1952-1954).

(iv) Any seasonal shifts of longitude of the troughs or ridges (Fig. 5 (b)) such as might imply lengthening or shortening wavelength appear rather ill-marked, apart from possible westward displacement of the broad Indian Ocean trough in summer to about 60°E. Tentative estimates both of the mean and modal wind direction at 3,000 and 6,000 m at Heard Island (53°S 74°E), based on seven months' radio-theodolite wind data in 1953 and supported by several years' experience of pilot-balloon data, were close to 270° in summer and close to 260° in the other seasons. Few, if any, other seasonal shifts of trough position can be detected from the evidence available. The prevailing direction shown by pilot balloons and cloud observations at medium- and high-cloud levels appears to be about 290° at all seasons at Macquarie Island (55°S 159°E).

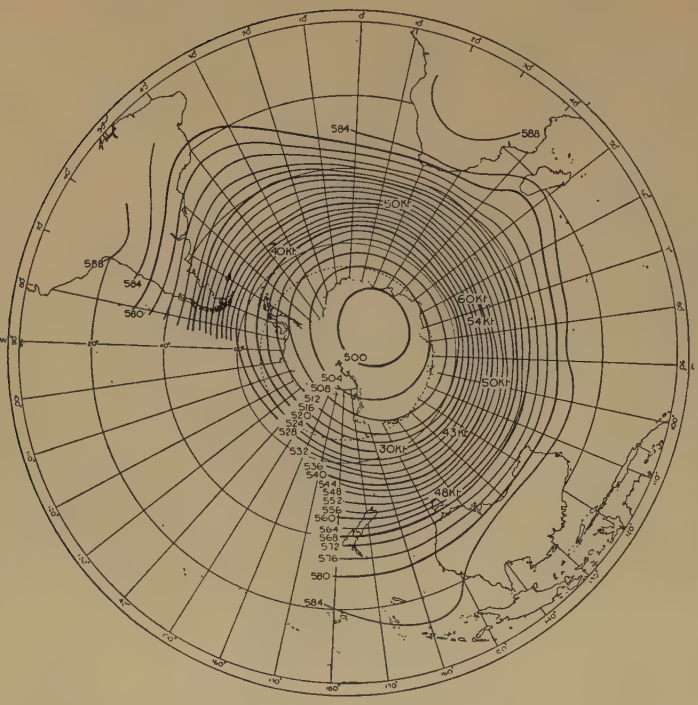


Figure 6. Mean 500 mb heights in dekametres, three summers, 1951-1954.

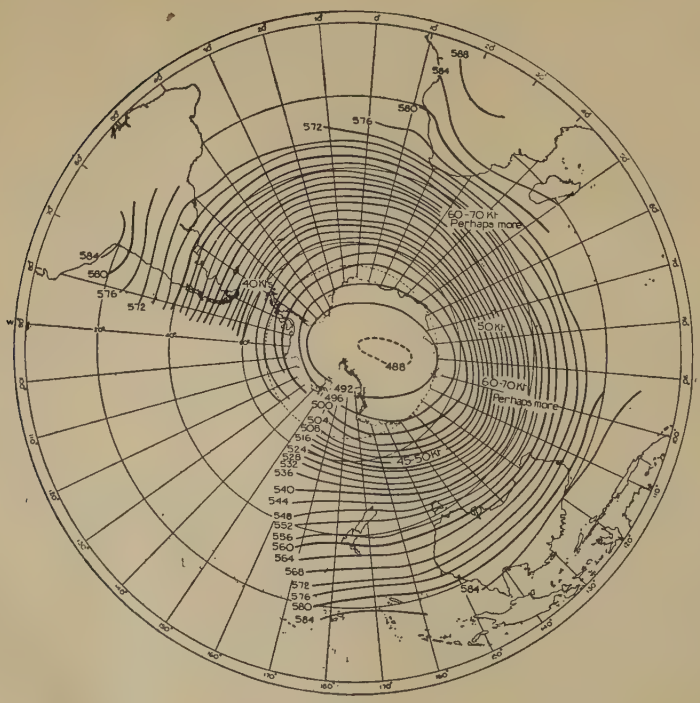


Figure 7. Mean 500 mb heights in dekametres, three winters, 1952-1954.

Any survey of troughs in the southern westerlies cannot be complete without mention of a trough in the Pacific sector, although data are extremely scanty. Two recent studies (Hofmeyr 1957, Rastorguev and Alvarez 1958) show a great trough, the presence of which could be definitely established in the first months of the International Geophysical Year (April-November 1957) because it sometimes came as far west as the Australasian sector. These were probably extreme positions associated with unusually meridional flow patterns. Hofmeyr shows a very broad trough, summer and winter, on average occupying the longitudes between 120° and 160°W. In some months this is probably the main trough in the hemisphere.

The rather surprising weakness of the mean trough in the southern part of the South Atlantic is confirmed by the resultant mean wind for the 36 months from the Stanley, Falkland Islands (51·7°S 57·9°W) radiosonde ascents, namely 264° 33kt. This figure fits well with Fig. 2 both as to direction and strength. The resultant mean winds at Stanley for the three summers, 259° 38 kt, and for the three winters, 265° 28 kt, fit equally well with Figs. 6 and 7. It is a peculiar feature of this part of the hemisphere that the 500 mb winds as shown by the maps and confirmed by the observed values are weaker in winter and spring than in summer and autumn. The summer maximum is believed to affect the winds at Stanley at all heights from the surface up to 200 mb; it is also seen in the surface winds at South Orkney (Signy Island). In some individual months at Stanley the mean 500 mb wind is reported to be from a point somewhat north of west, implying a ridge downstream from the longitude of the Andes.

The Andes, with peaks reaching 22,000 ft (6,500 to 7,000 m) at various latitudes between 9° and 33°S, are higher than the Rockies, but there are no peaks reaching 10,000 ft (3,000 m) south of 50°S. The New Zealand Alps exceed 12,000 ft (3,700 m) near 40°S, but their latitudinal extent is only about 5°. The main flow of the upper westerlies at 500 mb appears to pass south of both the Andes and the Southern Alps, though it is reported that the peaks of the Bolivian Andes near 23°S are at times swept by violent westerly storms associated with the high-level subtropical jet stream (Flohn 1955).

The trough in the 500 mb flow in the Indian Ocean sector cannot be attributed to any orographic barrier in the mainstream of the upper westerlies between 40° and 60°S. This trough is not only the biggest on the 3-yr mean map : it has also been judged to be the biggest on about 70 per cent of the individual monthly maps. It may be plausibly associated with the eccentric position of the broadest part of the Antarctic continent, the effective centre of whose cold radiating surface is displaced from the pole towards this sector (cf. Figs. 8 and 9).

With unusually well-marked surface southerlies in 50-70°S the cold air from East Antarctica is liable to be swept east into the Pacific cold trough, which must be thereby intensified in longitudes near New Zealand.

It is hard to avoid concluding that south of 45°S the dominant control of trough and ridge positions in the upper westerlies is thermal.

(c) *Polar outbreaks*

The paths of frequent outbreaks of polar air at the surface appear to fit the 3-yr mean 500 mb pattern (Fig. 2), and inasmuch as they have been known (for instance, to the whalers) for many years they tend to confirm the validity of this pattern over a much longer term of years.

Vowinckel (1956) shows the maximum frequency of polar outbreaks, revealed by analysis of 160,000 ships' observations near 64°S (period 1920-1955) as occurring over a broad sector centred about 110°E. Frequencies were only slightly lower in all longitudes between 60° and 150°E. Only a much narrower region between 60° and 35°W comprising the western and central parts of the Weddell Sea is believed to compare with the Indian Ocean – Australian sector.*

* The most northerly positions reached by occasional icebergs, under the combined influence of surface winds and water currents, are in the western South Atlantic and central Indian Ocean, in both cases about 35°-40°S.

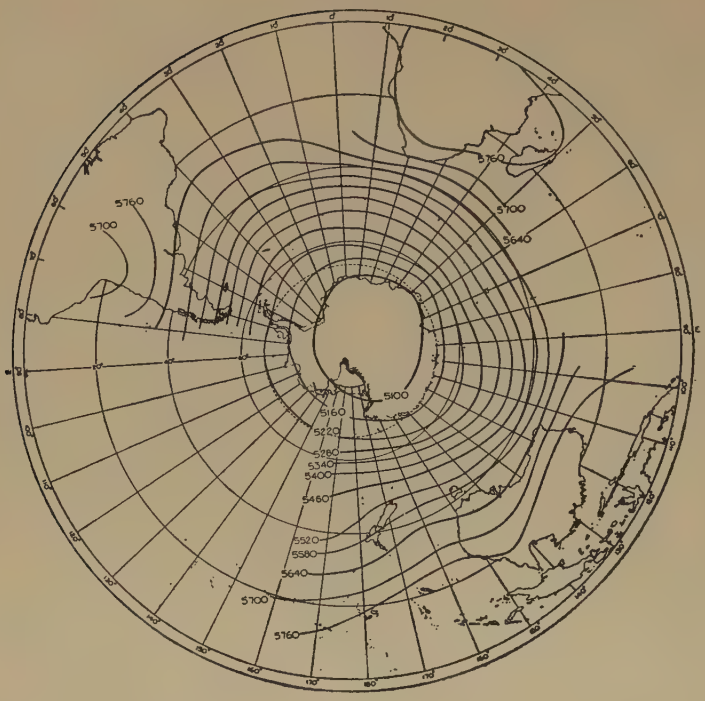


Figure 8. Mean 1,000-500 mb thickness in metres, three summers, 1951-1954.

Figure 9. Mean 1000-500 mb thickness in metres, three winters, 1952-1954.

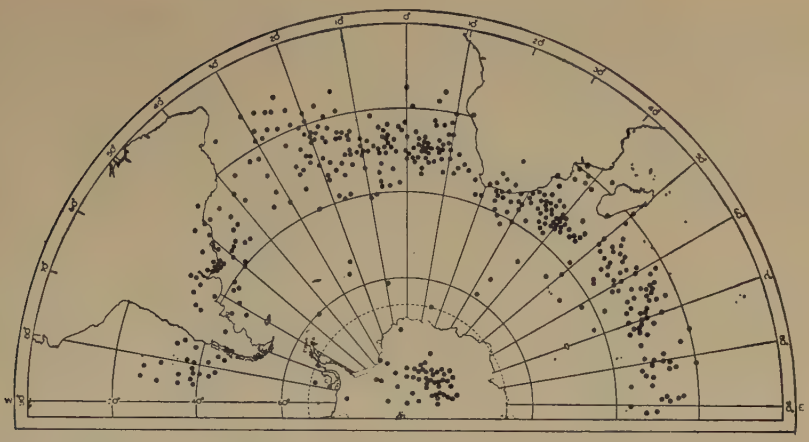


Figure 10. Distribution of once-daily positions of anticyclone centres, summer 1946-1947 (Dec.-Feb).

In parts of the Pacific sector and east of the Weddell Sea, surface winds with northerly components are known to be prominent.

Another preferred region for polar outbreaks just east of the northern tip of Enderby Land (66°S 52°E) has also been suspected, though this may be little more than a local effect superimposed on the general high frequency in the Indian Ocean sector.

The Weddell sea outbreaks sweep north and northeast to affect all parts of the South Atlantic (Britton 1949, Lamb 1957a) and sometimes turn east into the Indian Ocean.

Cold air from the South Polar Plateau and the vast interior of East Antarctica appears to flow out almost continuously over the Adélie Land coast and in a broad, generally east-southeasterly airstream over the fringe of Antarctica in the Indian Ocean sector between 150° and probably 50°E. This constancy is linked with the very great frequency of anticyclones centred near 80°S 60°E (cf. Figs. 10-12). The common southerly blizzards

TABLE 1. Frequencies (per cent) of various surface-wind directions at open sites in East Antarctica and at certain other points for comparison

Position and period	N	NE	E	SE	S	SW	W	NW	Calms
Mawson (67°S 62°E) (June-Oct. 1956)	1	2	11	59	9	3	2	0·5	13
Gauss (66°S 90°E) (12 months 1902-3)	0	2	57	11	5	3	9	2	11
'The Grottoes' Shackleton Ice Shelf (66°S 95°E) (10 months 1912-13)	0	0	35	48	3	2	4	1	4
Pionerskaya (70°S 95°E) (June-Oct. 1956, 2,700 m a.s.l.)	0·4	0·6	11	76	10	1	1	0·2	0
Port Martin, Adélie Land (67°S 141°E) (1950-51)	1	2	4	58	23	2	1	1	8
Snow Hill (64°S 57°W) (1½ years 1902-3)	3	13	5	1	15	38	7	1	17
Maudheim* (71°S 11°W) (22 months 1950-52)	0	14	52	4	18	9	2	0	–
Bay of Whales, Ross Ice Shelf (78°S 166°W) (1911-12, 1929-30, 1934-35, 1940-41. 4 years)	4	5	32	10	19	16	4	1	8

The sources of data used are listed at the end of this paper.

* The data for Maudheim are provisional, being based upon occasions when the windspeed was greater than or equal to 5 m sec^{-1}.

Figure 11. Distribution of anticyclone centres, summer 1951-1954. Percentage of synoptic maps with an anticyclone centre (defined by at least one closed isobar on a 5 mb spacing) within an area of 100,000 km² about any point.

Figure 12. Distribution of anticyclone centres, winter 1952-1954. Percentage of synoptic maps with an anticyclone centre (defined by at least one closed isobar on a 5 mb spacing) within an area of 100,000 km² about any point.

ear McMurdo Sound and in the western Ross Sea near 70°-75°S, 160°-170°E may be een as a branch of the same major outflow of cold air. Sample observational records o illustrate this quasi-permanent air current are summarized in Table 1. Snow Hill east side of Grahamland), Maudheim and Bay of Whales are included for comparison, o show the changes of dominant direction and rather greater scatter in other sectors.

The outflow of cold air from East Antarctica is only partly katabatic in origin, and liffers from the Greenland katabatic wind in giving rise to cold airmasses of great depth ver the ocean.

(*d*) *Intensity and breadth of the main belt of southern upper westerlies*

The overall range of yearly mean 500 mb height between the tropical and polar ones in the southern hemisphere (Fig. 2) materially exceeds the range in the northern emisphere. A greater total zonal circulation of the atmosphere over the southern emisphere is confirmed by Fig. 4.

The greatest yearly mean 500 mb heights are believed to be about 5,900 m (Flohn 955) over the high-level stations of the Bolivian Altiplano (La Paz, 17°S 68°W) and 5,870 o 5,880 m over the central African highlands. The nearest approach to these figures n the northern hemisphere appears to be in Mexico and the southwestern United States – Mirida (21°N 89°W) 5,883 m; Tacubaya (19°N 99°W) 5,866 m; Mazatlan (23°N 106°W) 5,852 m; Lake Charles (30°N 93°W) 5,824 m. It is probable (Flohn 1955) that the existence of a raised heating surface in low latitudes contributes to this result : raised surfaces of comparable extent in the northern hemisphere (e.g., Arizona and the Mexican ighlands, Spain, Anatolia, Persia, Afghanistan and Tibet) are mostly well north of 25°N nd get too cold in winter to produce such high annual values of the 500 mb surface.

Heights over Cocos Island (12°S 97°E) 5,870 m, and the equatorial zone of the outh Pacific near the Gilbert and Ellice Islands (0-10°S 170-180°E) 5,850 m also appear o exceed northern hemisphere equivalents : these places are remote from cold water currents and in or near the regions of warmest extensive ocean surfaces in the world see Riehl 1954, p. 37).

The lowest yearly mean 500 mb heights over Antarctica, 4,920-4,990 m, indicated by the *Notos* charts with support from radiosonde observations at Little America (1940-41), Maudheim (1950-52), and Port Martin (1951-52) are some 300 m below the lowest in he Arctic (cf. Fig. 4). Rather more than half of this disparity appears attributable to he prevalence of m.s.l. pressures 15-25 mb lower in the Antarctic than in the Arctic. There is no obvious disparity between prevailing pressures in the northern and southern subtropical high pressure belts; so the surface circulation, like that aloft, shows a greater otal of westerly momentum in the southern temperate zone than in the northern. The est of the disparity between the 500 mb heights over the Arctic and Antarctic is directly due to the colder Antarctic troposphere, especially in summer (cf. Figs. 8 and 9; see also Flohn 1952). Summer values of mean 1000-500 mb thickness over the Antarctic are 200-250 m lower than over the Arctic (south of 60°S precipitation normally falls as snow at all seasons). The winter thickness values are however similar in both northern and southern polar regions.

The steepest thermal gradient is in the Indian Ocean sector, between the trough in he higher latitudes and the particularly warm tropical zone of this ocean. This thermal gradient is confirmed by the fact that the islands in 50°-80°E exhibit climates ranging rom fully tropical with tropical vegetation at Mauritius (20°S) to ice age at Kerguelen 49°S) and Heard (53°S).

We should expect to find the strongest upper westerlies in this sector. This is nstanced by the extreme wind observation of 300 kt at 229 mb (11 km) on 27 Sept. 1956 at Amsterdam Island, 38°S 78°E (Schmitt 1956) and fits well with the distribution of average geostrophic speeds at 500 mb overprinted on Figs. 2, 6 and 7. The strongest 500 mb gradient in the northern hemisphere on Fig. 4 corresponds to a yearly resultant mean wind of about 42 kt over a restricted region of the Canadian maritime provinces

between 45° and 50°N. This figure is exceeded over most of a zone 1,200 to 1,500 mi wide between 40° and 60°S 20-150°E with peak values slightly over 50 kt in the annual mean and probably 60-70 kt or rather more in the means for June to October.

(e) *Polar front and subtropical jet streams*

The maximum winds indicated by the mean 500 mb map correspond to the polar front jet streams over the Southern Ocean. As in the northern hemisphere the subtropical jet stream is at higher levels. It is indicated at about 12 km on the various published vertical cross-sections (Loewe and Radok 1950, Flohn 1950, Hutchings 1951, Gibbs 1953, van Loon 1955), but these all suffer from lack of data south of 40°-45°S and do not show the maximum over the Southern Ocean associated with the polar-front jet streams at all well.

The greatest average westerly component in the subtropical jet stream probably exceeds 100 kt in winter in all the longitudes investigated. Over Australia and New Zealand its winter position is 25°-28°S; in summer it moves south, perhaps as far as 40-45°S and the maximum is thought to decline to 50-60 kt. The average latitude of this system over the eastern South Atlantic and western Indian Ocean is given rather further south than in the other sectors examined, namely at 35°-40°S in winter and 40°-45°S in summer, very much closer to the zone of the polar front jet streams than in the Australian sector.

The mean westerly components at 5-6 km indicated by the published cross-sections for the Atlantic and Indian Ocean sectors agree well with Figs. 6 and 7, both as to strength and position. Agreement between the 500 mb charts here illustrated and the published cross-sections in 150° and 170°E is less obvious, but this is probably because the strongest 500 mb winds in these longitudes are further south at 55°-65°S, too far south for the observations used on the cross-sections. Nevertheless all the cross-sections indicate a secondary maximum of wind speed somewhere south of 45°S.

These comparisons suggest that, as in the northern hemisphere, the subtropical jet stream has a somewhat separate existence from the jet streams of higher latitudes moving about in association with the polar front though at times coalescences may occur, especially in the Indian Ocean (Palmén 1954, Bannon 1954b, Lamb *et al.* 1957). The latitudes of the zone (40°-60°S) over which the mean winds at 500 mb are stronger than their northern hemisphere counterparts suggest that the disparity is principally associated with the polar-front jet streams of the Southern Ocean. The breadth of the belt with mean wind over 50 kt at 500 mb may be taken to imply great frequency of high wind speeds at this level in the polar front jet streams.

(f) *Seasonal variation*

The profiles in Fig. 5 (b) indicate a shortening of the wave spacing of troughs and ridges at 40°S from winter to summer in the South Atlantic and perhaps everywhere from the Andes to the Indian Ocean sector. The amplitude of the Indian Ocean trough is greatest in winter and spring : it becomes much less in summer,* though the great sweep of west-north-westerly winds between this and the western Pacific warm ridge is prominent at all seasons. The mean direction becomes almost north-west over the Ross Sea, as may be verified from Simpson's analysis (1919) of the frequency of different directions shown by the smoke from Erebus (4,000 m) and by medium and high cloud at Cape Adare and McMurdo Sound (cf. also Court 1949, Flohn 1952). No regular longitudinal shift of this north-westerly steering current with the march of the seasons can be identified from the evidence here used. Depressions are steered east and south-east in the Australian sector and into the Ross Sea at all seasons.

* During the southern summer, the Eurasian winter monsoon forces the inter-tropical convergence south to a mean January position along an arc from 15°S over Madagascar to 7°S between 80° and 100°E and 15°-18°S over north Australia. In approximately the same months the higher latitudes of the southern Indian Ocean are invaded by tropical storms, which, however, quickly occlude on their recurving tracks to the south-east.

The pronounced pressure minimum in late winter in the Ross Sea, culminating with ɔrmal October m.s.l. pressure about 973 mb (standard deviation *circa* 3 mb at Bay of Whales based on 9 years' data for areas in the southern Ross Sea about 78°S) is associated ith depressions steered by the upper north-westerly winds in the Australian sector. At ıe peak of this development, in September and October, low values of mean pressure ɛm regularly to spread across Antarctica (with the depression track) as far as southern ɾahamland and the Weddell Sea.

The summer and winter 500 mb maps (Figs. 6 and 7) and those for spring and ıtumn (not reproduced) indicate some general northward shift of the zone of maximum ind from summer to winter. Speeds increase in the lower latitudes in winter, whereas latitudes near 55-60°S the maximum is in summer as at Falkland (52°S). In the South merican sector the thermal gradient over quite a broad range of latitudes is actually ronger in summer than in winter. Meinardus (1947) has considered this latter phenomnon, chiefly in terms of surface pressure and temperature gradients, and found supporting ɣidence in the prevailing speeds of depressions : apparently the combined effect of ɛasonal shift of latitude of the strongest westerlies and rather small overall seasonal ıange of strength results in a fairly generally summer maximum of the surface winds 50°-60°S. There is, however, a general intensification of the flow around the broad ndian Ocean trough from summer through autumn to late winter and spring, followed y a rather rapid jump south and weakening of the maximum in early summer.

The progression is clearly seen near 70°E, where the maxima in the mean 500 mb flow ɔpear to be :

Summer	(Dec.-Feb.)	60 kt near 58°S
Autumn	(Mar.-May)	55 kt near 55°S
Winter	(June-Aug.)	60 kt near 50°S
Spring	(Sept.-Nov.)	65 kt near 50°S

Although the overall seasonal change of maximum speed is not great, there is a ɔteworthy broadening of the belt of strong flow over the whole Indian Ocean sector ɾom summer to late winter.

The seasonal movement appears to follow that of the ice limit, the maximum 500 mb ɾind lying some 8°N of the ice limit in this sector at each season.

The remaining sections of this paper are devoted to consideration of phenomena ɛlated to the southern westerlies, which may be used to provide verification of the ıaracteristics here described and demonstrate the apparent range of influence of the ɣstem.

3. Incidence of blocking anticyclones

In the northern hemisphere, investigators have found that when the upper westerlies ɛtain great strength over a considerable sector, large-amplitude waves usually develop ıd lead to quasi-stationary blocking patterns 2-3 days later with a great and rather ɛrsistent reduction of the zonal flow. Paradoxically, the mean speed of the 500 mb ɛsterlies in 40°-60°S for several months of the year over a very wide sector reaches ɑlues that would be high for the northern hemisphere, yet stationary blocking situations ɾe rare.

Figs. 10 to 12 show the distribution of once-daily positions of all surface anticyclones ɛfined by at least one closed isobar (5 mb spacing) in the summer of 1946-7 (individual ɔsitions, half hemisphere) and (by frequency isopleths) in the three summers 1951-4 ıd three winters 1952-4.* Similar maps have been plotted for all seasons of the year ıd for other years, but are not reproduced. Numbers printed on Figs. 13 (a) and 13 (b) ıow all cases south of 40°S in the summer and winter of 1952-3 where a stationary

* The consistency observed between the distribution indicated in the summers of 1946-7 (Fig. 10) and 1951-4 (Fig. 11) lends ɒport to the belief that the observation network, which was much sparser south of 40°S in the earlier year, has for some time been ɛquate for the purpose of this type of investigation (the Pacific sector was not covered by the daily analyses from which Fig. 10 ɜ compiled).

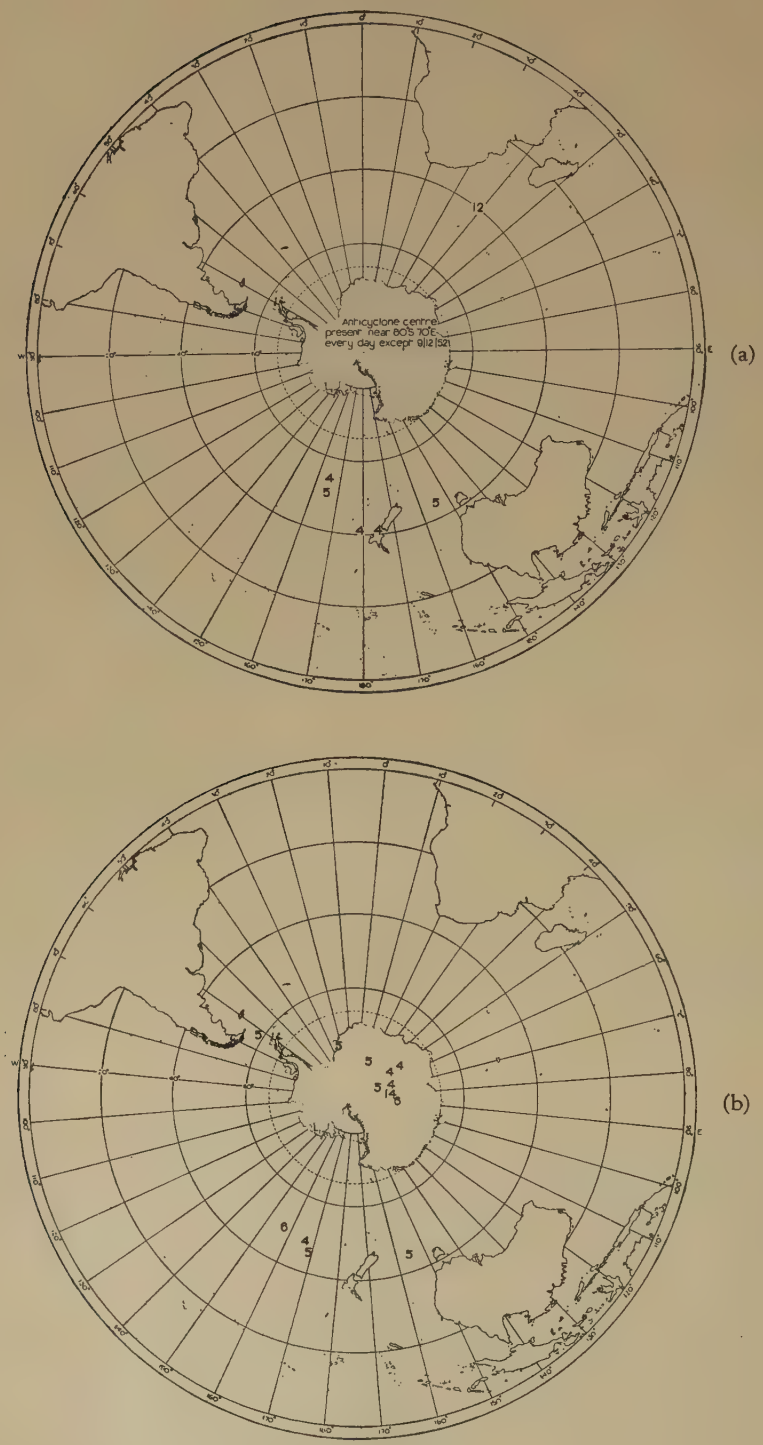


Figure 13. (a) Occurrences of quasi-stationary anticyclones south of 40°S, summer (DJF) 1952-1953. Centres which stayed x days within 600 naut mi of the position marked by x; x plotted in all cases where $x \geqslant 4$; (b) Occurrences of quasi-stationary anticyclones south of 40°S, winter (JJA) 1952. Centres which stayed x days within 600 naut mi. of position marked by x; x plotted in all cases where $x \geqslant 4$.

nticyclone remained for four days or more centred within 600 naut. mi of the same osition and record the number of days. It is found that quasi-stationary anticyclones outh of 40°S are limited to positions over Antarctica and three sectors of the surrounding cean, namely :

(i) From close to the south coast of Australia to a wider range of latitudes in the estern Pacific probably as far as the region near 60°S 120°-140°W.

(ii) Near the tip of South America and south-east thereof over the region of the cotia Sea in 60°-65°S; they are known to occur on rare occasions in longitudes farther ast in the South Atlantic sector 50°-60°S.

(iii) The region south-east of South Africa about the Marion and Crozet Islands n 40°-50°S in the west of the Indian Ocean sector.

Even in these regions the frequencies of stationary anticyclones do not appear to ompare with the corresponding zone of the northern hemisphere, and the vast regions vhere virtually no anticyclones occur (Figs. 11 and 12), whether stationary or mobile, re unmatched in the north.

The daily positions of anticyclones over the Southern Ocean south of 40°S show a reference for the same three sectors of the ocean. They almost entirely avoid the region f the great upper trough in the eastern part of the Indian Ocean sector, and are only eally common in the western-central Pacific sector.

As in the northern hemisphere, blocking shows a pronounced maximum in late vinter-early spring, but over the Southern Ocean the preferred positions are east, not vest of the major continents.

The three regions of principal occurrence of anticyclones over the ocean south of 0-50°S are in reasonable conformity with the upper-air trough and ridge positions evealed by the present investigation. They reflect :

(i) anticyclogenesis occurring to the warm side of the strong west-north-westerly ipper windstream in the Australasian sector well downstream from the Indian Ocean rough. The latitude tends to increase eastwards. In the individual cases the surface nticyclone is presumably east of an upper ridge developing and travelling east into the Pacific (cf. Schmitt 1957).

(ii) and (iii) anticyclogenetic patterns just west of the South Atlantic and Indian Ocean troughs.

The preference for the winter season is presumably linked with maximum amplitude f the troughs.

The Pacific sector, especially between 90° and 130°W, appears to be by far the most requent location of occasional great quasi-meridional (usually northwest-southeast) idges of high m.s.l. pressure linking the subtropical belt with the Antarctic. Simpson 1919) noted that surges of high pressure over Antarctica normally seemed to proceed rom this sector, though recent synoptic evidence demonstrates that surges sometimes enter East Antarctica from the South American sector by way of the Weddell Sea and more ccasionally directly from the ocean in 0°-60°E.

Blocking patterns do occur fairly frequently over certain parts of the Southern Ocean, nd there is no reason to exclude the possibility that they follow periods of excessive zonal low (' hydraulic jump '); but stationary situations seldom result. The anticyclones over he ocean always tend to drift away east, no doubt because of the great overall thermal radient in the zone 40°-60°S. Cells do, however, split off over Antarctica, where there is ittle thermal gradient, and become stationary there. This process is capable of renewing he Antarctic anticyclone after a period of low pressure in high latitudes.

Apparently, in the northern hemisphere the great continents in the temperate zone lay an important part in permitting stationary patterns to develop, thanks to wide areas f slight thermal gradient often associated with the main winter snow cover, and the

development of distorted thermal patterns between oceans and continents in the same latitude zone. In the southern hemisphere geography does not generally permit large depressions or anticyclones to produce a thermal pattern capable of maintaining the system in a stationary position.

4. Depression tracks

Our knowledge of the commonest depression tracks over the Southern Ocean, supported by the observed distribution of glaciation described in the next section, provides further confirmation of the general upper air pattern south of 40°S and implies that the Indian Ocean trough has been dominant for a very long time.

A sample map of the distribution of frequencies of depressions south of 30°S, originally published elsewhere (Lamb and Britton 1955), is reproduced in Fig. 14. The main frequency maxima are (i) near 65°S 20°E south-east of the South Atlantic trough, (ii) at 62°-65°S in the Indian Ocean sector between 80° and 120°E, (iii) along a broad depression path from near 50°S in the Australian sector into the south-eastern part of the Ross Sea and (iv) in the Bellingshausen Sea near 65°S 70°W. These positions show the expected associations with the troughs in the 500 mb flow (Figs. 2, 6, 7). Depression frequencies all around the zone 60°-70°S are kept high by the general, though often rather slow, eastward drift of occluded and decaying centres. Occasionally, in many longitudes, these centres are swung south and west on looped tracks over the fringe of Antarctica under the influence of the circulations of more vigorous younger depressions over the ocean to the north. The map also indicates several favoured paths and origins for frontal waves

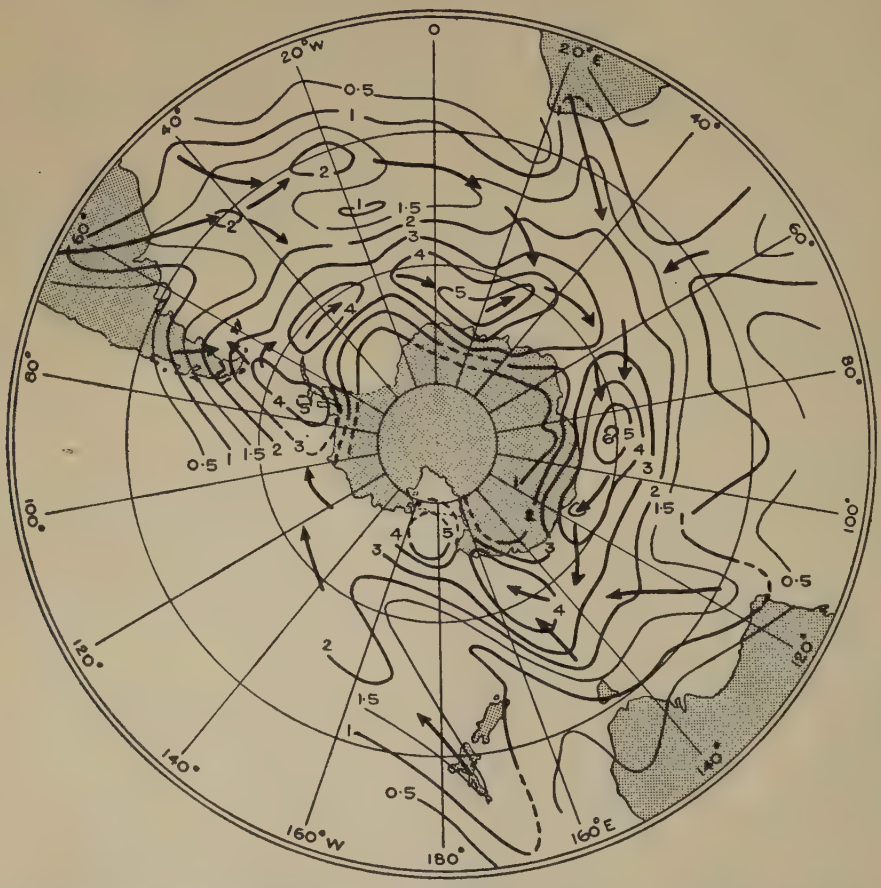


Figure 14. Depression frequencies; Jan.-Feb. 1947; mid-Mar.-mid-April 1947; mid-Mar.-mid-April 1948. Arrows indicate the general direction of movement.
(Divide the figures by 2·3 to obtain percentage of synoptic maps with a depression centre within 100,000 km^2 of any point).

isturbances. Data available consistently indicate the lowest mean annual pressure at n.s.l. near 60°-65°S in the eastern part of the Indian Ocean sector in 90°-110°E where the alue is possibly as low as 980 mb.

In spite of the severe shortage of observations for the Pacific sector, there seems eneral agreement that a mean surface pressure trough exists with a long northwest-outheast axis between the southern anticyclones in 40-60°S in the sector New Zealand o 140°W and the quasi-permanent Easter Island anticyclones near 30°S in the eastern outh Pacific. This corresponds to a depression track (Fig. 14) from the cyclogenetic egion in 20°-30°S between New Zealand and Tahiti to Drake Passage and the Bellings-ausen Sea in 55°-70°S 60°-80°W (cf. Bjerknes *et al.* 1933, Shaw 1936, Gibbs 1953, Lamb 956, Mansfield and Glassey 1957, Schmitt 1957). Southerly wind components at the ear of these depressions are the likely explanation of the northward bulge in the frequency f icebergs (fairly commonly reaching 50°S, occasionally 45°S) in the central Pacific etween 105° and 135°W (Shaw 1936) and support the existence of a trough in the upper vesterlies.

5. Pattern of glaciation

Glaciation of land masses depends on sufficient precipitation and temperatures emaining near or below 0°C. In middle and high latitudes adequate precipitation requires yclonic activity.

There are a number of islands in about the same latitude and in some cases of similar elief, which enable us to make comparisons between conditions in different longitudes.

The most northerly island in the southern hemisphere with many great glaciers some f which reach the sea is Kerguelen, 70 mi long near 49°S 70°E, whose summit reaches ver 6,000 ft (1,865 m). Heard Island, 25 mi long near 53°S 74°E with its summit reaching ,000 ft (2,750 m), is completely glaciated but for a few small ice-free areas on the lower lopes. Bouvet Island (54°S 3°E) is a volcanic cone, 5 mi across at sea level and reaching ver 3,000 ft (935 m) : the summit and the steep northern and western sides are ice-free ut often snow-covered; the whole eastern side is under an ice sheet. South Georgia is much bigger and very mountainous island, 100 mi long and 20 mi broad near 54°S 38°W nd its highest summit reaches to about 9,300 ft (2,840 m). There is much ice amongst the nountains in South Georgia and some glaciers reach sea level, but the permanent snow ine is generally about 2,000 ft, falling to 1,000 ft (300 m) at the north-western end where laciation is relatively heaviest. Macquarie Island, near 55°S 159°E, 18 mi long and ummit about 1,400 ft (430 m), loses all its snow cover in summer. These islands, all bout 50°-55°S, support the diagnosis that a great, broad, quasi-permanent upper cold rough occupies the Indian Ocean sector, whereas South Georgia and Macquarie Island enefit to some extent from warm ridges, respectively some distance west and east of them.

The great ice-sheet of East Antarctica derives nourishment from snowfall in cyclonic ituations all along its fringe, and there is evidence to suggest that depressions and frontal roughs pass with fair frequency across the sector 100°-170°E, perhaps mainly on recurving racks from the Ross Sea. However, there is little doubt that the most frequent depressions ver Antarctica are systems travelling south-east into the Ross Sea and proceeding to cross art of West Antarctica, occasionally right to the Bellingshausen and Weddell Seas. For he present, it remains a mystery why West Antarctica, which is very mountainous, with highest reported peak 20,000 ft (6,000 m), appears much less heavily glaciated than East Antarctica; the ice summits are believed to be about 5,000-7,000 ft (2,000 m) and 14,000 ft 4,000 m) respectively near 75°-80°S 100°-120°W and 80°S 50°E (Sharp 1956). The de-ression tracks observed at the present epoch would suggest increasing glaciation in West Antarctica and possibly insufficient precipitation to maintain full nourishment of the East Antarctica ice dome.

Continental regions exceeding 1,500 naut. mi in longitudinal extent in high latitudes ppear to have a critical importance in establishing a fixed cold trough in the upper-air low (Lamb 1955). A cold trough should therefore be sooner and more readily established ver East Antarctica than over West Antarctica, which is scarcely 1,000 mi across.

6. World-wide relationships : seasonal and secular changes

There is at all seasons an obvious deficit of atmospheric mass everywhere south of about 40°S in comparison with the corresponding parts of the northern hemisphere. Over much of the region this disparity is of the order of 20 mb.

Calculations based on the best available world maps of average pressure at m.s.l. suggest that the deficiency, as compared with the northern hemisphere, begins about latitude 45° in the southern winter but extends to all latitudes in the southern summer. Over the year there is an apparent excess of mass over the northern hemisphere as a whole of about 0·5 per cent, equivalent to about 5 mb of average m.s.l. pressure; the excess appears to range from about 0·9 per cent in the northern winter to 0·1 per cent in the northern summer.

When allowance is made for the intrusion of elevated land in either hemisphere, the apparent difference of one half per cent of atmospheric mass between northern and southern hemispheres, is approximately cancelled out. (Much estimation is involved, but the correction is clearly of the same order and seems to be close to the same value). There is no doubt however of a disparity of over one per cent between latitudes south of 35°S and north of 35°N. Between 5° and 30°S the annual figures suggest a small excess over northern hemisphere equivalents. This excess is related to the zone of subtropical maximum pressure in the southern hemisphere being about 5° nearer the equator than its northern counterpart when either respective winter or summer positions are compared.

Various estimations of the mean latitude of the meteorological equator (intertropical convergence) based on latitudes of the subtropical anticyclone centres, limits given by various authors for the doldrums belt in different sectors, penetration of the northeast and southeast Trade Winds and reported crossing of the geographical equator by old cold fronts, principally of southern origin, give corroborative figures about 3·5°N for the overall mean latitude of the intertropical convergence. Only in the western Pacific sector, west of about 160°W, is the annual mean position south of the geographical equator (about 3°S): it is perhaps significant that these are about the same longitudes as those of the main warm ridge in the southern upper westerlies.

It is not so easy on geographical grounds to make a direct comparison of the respective positions of the sub-Arctic and sub-Antarctic zones of minimum pressure. A mean latitude 60°-65°S may be compared with rather north of 65°N for the northern minimum, which is, however, strongly affected by the land distribution in its latitude zone.

The average northward displacement of all the climatic zones* in spite of the deficit of atmospheric mass to the south of them, may appear paradoxical. Both phenomena must however be seen as adjustments to (i) the latitudinal position of the main thermal gradient around the respective polar caps, and (ii) the outstanding momentum and energy of the southern westerlies.

The momentum and energy of the southern westerlies appear fundamental and must be associated with the lesser friction over the southern hemisphere.

A balance is attained at a mean latitude somewhat north of the equator between the centrifugal tendencies of the planetary circulations over the northern and southern hemispheres. The meridional circulation of the southern hemisphere extends beyond the geographical equator. This is in keeping, too, with the fact noted by Faust (1955) and others (using observations over Europe), that a net ageostrophic transport of air towards the warm side of the jet stream tends to be developed, increasing with the intensity of the maximum wind.

Comparisons of the monthly mean pressure over the southern polar cap with the varying latitude and intensity of the westerlies suggest certain associations. Mean m.s.l. pressures over the polar cap (70°-90°S) and in the sub-polar zone 60°-70°S were compared with m.s.l. pressure, 1,000-500 mb thickness and 500 mb heights and gradients in the zone of westerlies over the best covered longitudes 70°W-170°E from monthly data published

* The latitudes of the ice-covered islands in the Southern Ocean mentioned on page 19 should also be compared with northern equivalents. In Chile the northernmost glacier (San Rafael) from the Patagonian ice-cap reaches the sea in 46° 40'S, ten degrees farther from the Pole than any Alaskan glacier.

Notos over periods of up to 41 months. Higher than average pressure south of 60°S as found, as would be expected, highly significantly correlated with weakened surface essure gradients for westerly winds in all latitudes south of 45°S ($r = -0{\cdot}67$ to $-0{\cdot}77$ r 29-41 months) and with equatorward displacement of the latitude of strongest surface esterlies ($r = -0{\cdot}73$ for 29 months). Relatively high pressure south of 60°S was gnificantly correlated with below-average values of 1,000-500 mb thickness at 60°S $= -0{\cdot}50$ to $-0{\cdot}65$ for 27 months) and showed weaker correlations with lowered ickness also at 50° and 40°S : there was a suggested, though inconclusively demonstrated, sociation with increased thickness gradient between 40° and 60°S ($r = +0{\cdot}14$ to $+0{\cdot}35$). here was a suggested association of high pressure everywhere south of 60°S with weak- ied westerly flow at 500 mb in 50°-60°S ($r = -0{\cdot}11$ to $-0{\cdot}24$ over 32 months), equator- ard shift of the latitude of strongest flow at this level ($r = -0{\cdot}27$ to $-0{\cdot}33$ over 31 onths) and strengthened flow in 30°-40°S ($r = +0{\cdot}23$ to $0{\cdot}40$ for 32 months), but none these relationships was conclusively demonstrated. The most definite upper-air rrelations were with pressures in the subpolar zone rather than south of 70°S, where .s.l. pressures are moreover largely fictitious and uncertain.

Both seasonal and secular climatic trends associated with general strengthening or eakening of the zonal circulation and of the meridional temperature gradients affecting ie position or intensity of the southern westerlies should have unlike effects south and orth of this system and entail appreciable adjustments over the rest of the world.

Many years ago Meinardus (1909) noticed that south of about 50°S the highest m.s.l. ressures appeared to occur in summer (December-January), in spite of the complication a double wave in the annual pressure curve in 60°-70°S in the Weddell Sea-Atlantic ector with another maximum in May or June (Schwerdtfeger and Prohaska 1956). The anti-monsoonal' trend appears dominant in the highest southern latitudes, and about 8°S in the Ross Sea the summer maximum appears to be the only significant one; winter iere is a season of low pressure, especially late winter. Meinardus made some calcula- ons of the seasonal exchange of atmospheric mass involved. The null point in this xchange appeared to be approximately the latitude in which the strongest upper westerlies ssociated with the polar-front jet streams are now seen to lie.

Willett (1950) and others have noticed unlike climatic trends south of about 40°S nd over the rest of the world. During the period of great warming of the Arctic, for istance, culminating between 1900 and about 1938, latitudes south of 40°S are believed y Willett to have been rather colder than formerly, and Kraus has suggested (1954) that ie westerly disturbances of the Southern Ocean were displaced north to affect Australia iore in approximately those years than before or since. Since about 1945 there are signs Vowinckel 1956) of an upward trend in the surface temperatures over the Southern)cean south of 60°S and, since about the 1920's, of depression tracks keeping gradually rther south in the Australasian sector (de Lisle 1957). The latest trend appears to co- icide with a levelling-off of the warming trend over Spitsbergen and northern Europe, ossibly the substitution of a rather colder régime with depression tracks somewhat rther south.

There are perhaps indications of opposite climatic trend over high southern latitudes om the rest of the world over a rather longer period : the farthest south positions eached by Cook near 40°E in 1773 and near 110°W in 1774 as well as by Bellingshausen 1820 near the Greenwich meridian and Biscoe in 1831 between 0° and 30°E were all °-2°S of the normal limit of pack-ice in recent years at the end of the melting season Macintosh and Herdman 1940), and in 1823 Weddell found the Weddell Sea almost free ice – an experience never repeated since. When the frailty of the wooden ships of iose days is remembered, it seems that the ice limit tended to be some degrees south of s supposed present normal position (as appears to be the view of the Scott Polar Research istitute). Herdman, however, doubts whether the modern normal is sufficiently deter- inate for this to be settled. There is nothing surprising in this since year-to-year ariations are considerable and are liable to exceed in any particular sector the supposed iange of long-period average position since the epoch 1770-1830.

The same overall trend might be deduced from the present advanced state of the South Georgia glaciers, which are generally in contact with their side moraines down to sea level and are known to have advanced since the 1880's, but show erratic changes in the past 6-12 years. There is apparently no bare rock which has not obviously been exposed to the weather for decades or centuries.

Between 1770 and about 1840 the ice limit in the North Atlantic was near Iceland for much of the year, and it was a cold epoch all over northern Europe (Koch 1945). The westerly winds on the North Atlantic and over western Europe were less strong, or less prevalent, than now and the climate zones in this sector of the northern hemisphere were certainly displaced some degrees south from the present-day experience.

Superimposed upon any tendency for sympathetic shifting of all the climatic zones, it may be that there are periods of general weakening or intensification of the circulation, intensification being commonly held to entail a tendency for equatorward displacement of the main upper westerlies in both hemispheres. Kraus (1954, 1955a, 1955b, 1956) has postulated general intensification of the upper westerlies in both hemispheres from about 1895 to the 1920's or later, causing general intensification of the subtropical anticyclones and narrowing of the equatorial rain-belt.

It seems that changes in the southern westerlies might affect the general atmospheric circulation over the rest of the world in at least three ways :

(i) changes of overall intensity of the circumpolar whirl should be directly related to changes in intensity of the subtropical anticyclones in the southern hemisphere and therewith of the southeast Trade Winds. Such changes bring about changes in the supply of warm water to the Gulf Stream and in the value of Walker's 'southern oscillation,' which involves changes of mean pressure over much of the world (it may be noted that variations of average m.s.l. pressure south of 50°S would have about ten times the magnitude of any general compensatory changes of pressure level, if spread evenly over the rest of the world).

(ii) changes in the mean latitude of the circumpolar whirl in the southern hemisphere are likely to be accompanied sooner or later by small, but probably appreciable, sympathetic shifts of other climatic zones.

(iii) changes of eccentricity of the southern whirl occur over short periods but any long-term changes might imply or require changes in the relative states of ice-cover over East and West Antarctica. These can probably be ruled out within the period of our meteorological instrumental record, but might be important in considering the ice ages and epochs lasting several millennia.

7. Acknowledgment

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Long waves, lee waves and gravity waves

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SUMMARY

It is shown that the equation determining the streamlines of large-scale atmospheric motions (jet streams and long waves) is similar to the equation determining the streamlines of small-scale gravitational motions of a stratified liquid. Both motions are characterized by the occurrence of jets, the number of jets being determined by the value of a dimensionless parameter. Atmospheric flow at 500 mb is compared with theoretical solutions (derived by Long 1955) of the non-linear equations for the flow of a stratified liquid. Linearization of the equation for large-scale atmospheric motions leads to the Rossby wave equation, whereas linearization of the equation for the motion of a stratified liquid leads to Scorer's lee-wave equation.

1. INTRODUCTION

It is a remarkable fact that the equations which govern the steady large-scale motion of the atmosphere are almost identical with the equations for small-scale gravity oscillations of a stratified liquid. Consequently laboratory observations on a small scale in a *non-rotating system* can be used to indicate possible steady motions of long waves and jet streams. Such laboratory observations have already been published, notably by Long (1955), though the direct application to large-scale motions does not seem to have been realised. The motions are characterized by the occurrence of jets, the simplest cases having a single meandering jet and the more complicated cases having several jets splitting and recombining to form a complex pattern. The number of jets is determined by the modified Froude number for stratified flow and by an analogous dimensionless number for large-scale atmospheric flow.

2. NOTATION

For the sake of brevity the large-scale atmospheric motions will be called ' long waves ' and the oscillations of a stratified liquid will be called ' gravity waves.' In the case of long waves the x axis is taken west-east and the y axis poleward ; in the case of gravity waves the x axis is horizontal and the y axis vertical. Equations with the letter ' a ' following the equation number refer to long waves, the letter ' b ' refers to gravity waves. The symbols used throughout the paper are listed below :

c speed of sound
f Coriolis parameter
F Froude number
g gravitational acceleration
h distance of a streamline from the x axis
L a characteristic length
p pressure
s logarithm of the density
t time
u x component of velocity
v y component of velocity
U a characteristic velocity
V speed $(u^2 + v^2)^{\frac{1}{2}}$

β $\frac{df}{dy}$

η vorticity $\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$

ρ density

ψ stream function

3. The vorticity equations

The following assumptions are made in the derivation of the equations :

(i) the motion is steady.

(ii) the motion is confined to the x, y planes.

(iii) the fluid is incompressible and frictionless.

(iv) the earth is assumed flat for long waves except for the variation of f with latitude.

The incompressibility condition does not imply absolute constancy of the density ut merely constancy along the trajectories (which in this case coincide with the streamlines) hence

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = 0 \quad . \quad . \quad . \quad . \quad . \quad (1)$$

$$u\frac{\partial \rho}{\partial x} + v\frac{\partial \rho}{\partial y} = 0 \quad . \quad . \quad . \quad . \quad . \quad (2)$$

In the case of long waves it is usual to assume that the density is an absolute constant; his assumption is not made here. Eq. (1) implies the existence of a stream function ψ uch that

$$u = -\frac{\partial \psi}{\partial y} \quad . \quad . \quad . \quad . \quad . \quad . \quad (3)$$

$$v = \frac{\partial \psi}{\partial x} \quad . \quad . \quad . \quad . \quad . \quad . \quad (4$$

Eq. (2) indicates that the density is a function of ψ only and consequently that $d\rho/d\psi$ a function of ψ only. Whence $d\rho/d\psi$ is constant along a streamline, viz.,

$$\frac{d\rho}{d\psi} = \frac{1}{v}\frac{\partial \rho}{\partial x} = -\frac{1}{u}\frac{\partial \rho}{\partial y} \quad . \quad . \quad . \quad . \quad (5)$$

nd

$$u\frac{\partial}{\partial x}\left(\frac{d\rho}{d\psi}\right) + v\frac{\partial}{\partial y}\left(\frac{d\rho}{d\psi}\right) = 0 \quad . \quad . \quad . \quad . \quad (6)$$

The equations of motion are :

$$u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} - vf + \frac{1}{\rho}\frac{\partial p}{\partial x} = 0 \quad . \quad . \quad . \quad (7a)$$

$$u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + \frac{1}{\rho}\frac{\partial p}{\partial x} = 0 \quad . \quad . \quad . \quad (7b)$$

$$u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + uf + \frac{1}{\rho}\frac{\partial p}{\partial y} = 0 \quad . \quad . \quad . \quad (8a)$$

$$u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + g + \frac{1}{\rho}\frac{\partial p}{\partial y} = 0 \quad . \quad . \quad . \quad (8b)$$

Straightforward manipulation of these equations leads to the generalized vorticity equations :

$$\frac{D}{Dt}\left(\eta + \beta h + \frac{ds}{d\psi}\tfrac{1}{2}V^2\right) = 0 \quad . \quad . \quad . \quad (9a)$$

$$\frac{D}{Dt}\left(\eta + g\frac{ds}{d\psi}h + \frac{ds}{d\psi}\tfrac{1}{2}V^2\right) = 0 \quad . \quad . \quad . \quad (9b)$$

showing in each case that the generalized vorticity is constant along a streamline. Eq. (9a) is the standard equation for the conservation of absolute vertical vorticity modified to take account of the variation of density across the flow. The term $\frac{D}{Dt}\left(\frac{ds}{d\psi}\frac{1}{2}V^2\right)$ can also be written in the familiar solenoid form, viz., $\frac{1}{\rho^2}\left(\frac{\partial\rho}{\partial y}\frac{\partial p}{\partial x} - \frac{\partial\rho}{\partial x}\frac{\partial p}{\partial y}\right)$. However, this form is not as suited to present purposes as the conservation form (9a). The effect of density variations is negligible except where v is small and the solenoids are dense. Eq. (9b) is the horizontal vorticity equation developed by Long (1953) and solved for certain types of boundary conditions (Long 1955). *These equations are identical in form, β corresponding to $g\,ds/d\psi$.* The variables can all be put in terms of the stream function and consequently Eqs. (9) completely determine the streamline patterns and velocity fields provided appropriate boundary conditions are given.

4. Similarity conditions

If the two flow patterns represented by Eqs. (9) are to be similar, the boundary conditions and scales of motion must satisfy certain relationships. The relationships can be determined by putting the equations in dimensionless form. Let U be a characteristic velocity and L a characteristic length, suffixes a and b denoting long waves and gravity waves respectively, then if asterisks signify dimensionless variables defined by :

$$\eta^* = L\eta/U, \qquad h^* = h/L, \qquad V^* = V/U, \qquad \psi^* = \psi/UL$$

the equations become :

$$\frac{D}{Dt}\left(\eta^* + \frac{\beta L_a^{\,2}}{U_a}h^* + \frac{ds}{d\psi^*}\tfrac{1}{2}V^{*2}\right) = 0 \quad . \quad . \quad (10a)$$

$$\frac{D}{Dt}\left(\eta^* + \frac{gL_b}{U_b^{\,2}}\frac{ds}{d\psi^*}h^* + \frac{ds}{d\psi^*}\tfrac{1}{2}V^{*2}\right) = 0 \quad . \quad . \quad (10b)$$

For simplicity consider a region between two straight streamlines parallel to the x axis, a distance L apart. If similar boundary conditions are given on a line intersecting the flow, i.e., similar vorticity, velocity and density distributions, then the flow patterns will be identical provided :

$$\frac{\beta L_a^{\,2}}{U_a} = \frac{gL_b}{U_b^{\,2}}\frac{ds}{d\psi^*} \quad . \quad . \quad . \quad . \quad . \quad (11)$$

This condition cannot usually be satisfied exactly since β is a function of y only, and $\frac{ds}{d\psi}$ is a function of ψ and therefore of y and x when the flow is not horizontal. We will assume with sufficient accuracy for present purposes that both are constant. On this assumption $ds/d\psi^*$ is equal to $\Delta\rho/\bar{\rho}$ where $\Delta\rho$ is the difference between the density at the bottom and the density at the top of the flow and $\bar{\rho}$ is a mean density (such that the total mass flow between the straight streamlines is $\bar{\rho}UL$) whence :

$$\frac{U_a}{\beta L_a^{\,2}} = F^2 = \frac{U_b^{\,2}\,\bar{\rho}}{gL_b\,\Delta\rho} \quad . \quad . \quad . \quad . \quad (12)$$

The right-hand side of Eq. (12) is the modified Froude number for stratified flow its square, depending on the definition adopted). Consequently the left-hand side can regarded as the counterpart of the Froude number for large-scale atmospheric motions. is number also appears in Rossby's theory of ' blocks ' where it plays the same part the Froude number in the theory of hydraulic jumps (Rossby 1950).

In many flows of interest it is found that

$$\frac{D\eta}{Dt} \gg \frac{D}{Dt}\left(\frac{ds}{d\psi}\tfrac{1}{2}V^2\right) \quad . \quad . \quad . \quad . \quad . \quad (13)$$

l the equations reduce to

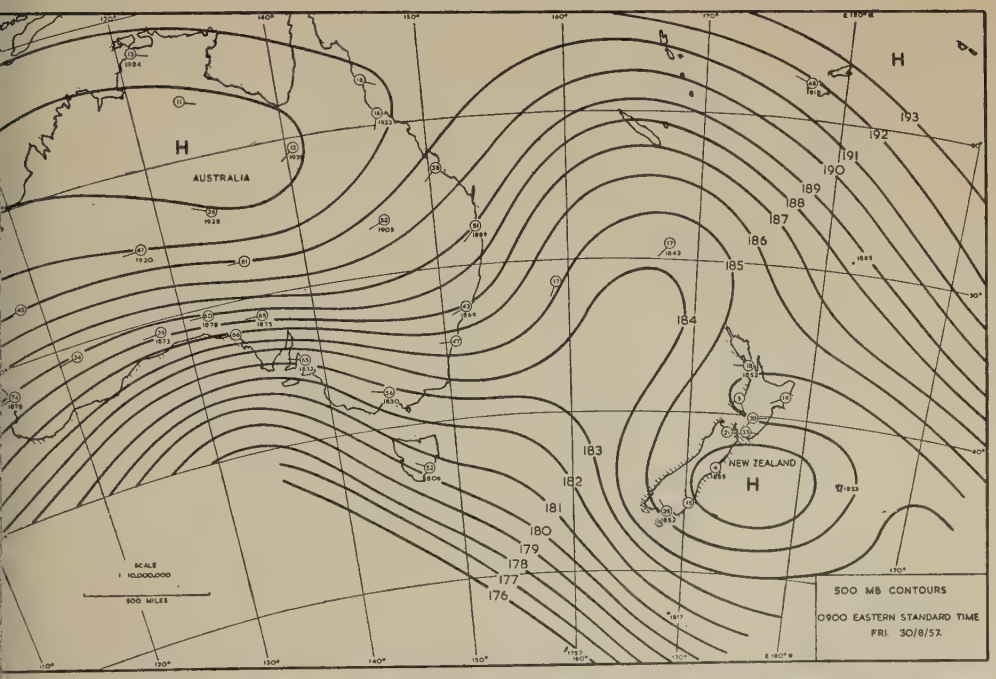
$$\frac{D}{Dt}(\eta + \beta h) = 0 \quad . \quad . \quad . \quad . \quad (14a)$$

$$\frac{D}{Dt}\left(\eta + g\frac{ds}{d\psi}h\right) = 0 \quad . \quad . \quad . \quad . \quad (14b)$$

This will be the case if the density variations are small, as they often are. The type approximation involved is the Rayleigh or Boussinesq assumption that the density iations can be neglected except in so far as they modify the gravitational attraction per it volume, i.e., the inertia of the liquid is assumed to be unaffected by the density iations. Under these circumstances it is unnecessary for the density distributions to be nilar along the boundary, similarity of the velocity and vorticity distributions being all at is required.

According to Long the ranges of values of the Froude number corresponding to ious patterns are :

$F > 1/\pi$	no jets
$1/\pi > F > 1/2\pi$	single meandering jet
$1/2\pi > F > 1/3\pi$	alternation between a double and a single jet
$1/n\pi > F > 1/(n+1)\pi$	alternation between $n-1$ jets and n jets.



ure 1. Double-jet situation over the Tasman Sea, 30 August 1957. The 500 mb contours are at intervals 100 ft. Numbers inside the circles give the wind speed in knots. Numbers outside the circles give the height in hundreds of feet.

In order to see whether Long's experiments and calculations have a direct larg scale application, consider the double-jet situation over the Tasman Sea on 30 Augu 1957 (Fig. 1). The corresponding Froude number is difficult to determine exactly sin the motion is not confined between two straight streamlines and the southern part of t system is not covered by the observations. However, its rough value is given by takin

$$L_a = 4000 \text{ km}$$

$$U_a = 5 \text{ m sec}^{-1}$$

$$\beta = 2 \times 10^{-11} \text{ metres}^{-1} \text{ sec}^{-1}$$

whence $$F^2 = U_a/L_a^2 \beta = 0{\cdot}0156$$

and $$F = 0{\cdot}125$$

U_a is the difference between the mean zonal speed (about 17 m sec^{-1}) and the speed movement of the system (about 12 m sec^{-1}).

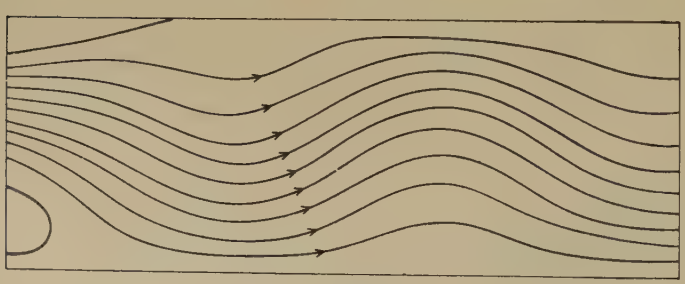
Flows with values of F of 0·200 and 0·143 are shown in Figs. 8 and 9 respectively Long's (1955) paper. The theoretical patterns from these figures are reproduced in Fig. of this paper. The atmospheric value for the synoptic situation considered here li within the Froude number range for a pattern alternating between a single and a double je

5. The pressure field

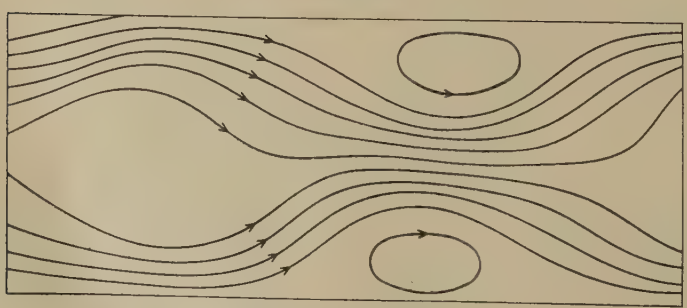
It has been shown in the preceding sections that under certain circumstances t streamline patterns of long waves are similar to those of gravity waves. It will now b shown that the pressure fields are *not* similar. This can most easily be seen by considera tion of the Bernoulli equations for the two types of motion. These are :

$$\frac{D}{Dt}\left(\tfrac{1}{2}V^2 + \frac{p}{\rho}\right) = 0 \quad . \quad . \quad . \quad (15a)$$

$$\frac{D}{Dt}\left(\tfrac{1}{2}V^2 + \frac{p}{\rho} + gh\right) = 0 \quad . \quad . \quad . \quad (15b)$$



(A) SINGLE MEANDERING JET



(B) DOUBLE JET SYSTEM

Figure 2. Theoretical streamline patterns from Long (1955).

It is immediately apparent from these equations that if the velocity fields are similar hen the pressure fields will differ by an amount dependent on gh. In fact the pressure eld for long waves is similar to the deviation of the pressure from its hydrostatic value gravity waves.

It seems desirable to dispense with the rather unrealistic straight streamline boundaries far considered and examine flow bounded by free streamlines. Consider an atmospheric t stream flowing between regions of constant pressure and stagnant air. The free treamlines along the boundaries of the jet are characterized by a constant pressure and onsequently, from Eq. (15a) the speed will also be constant. To produce a similar istribution of velocity in a stratified liquid we must have $p/\rho + gh$ constant along the ounding free streamlines. This can be arranged by having a stream of variable density owing between two layers of stationary liquid of constant density (see Fig. 3), the density eing continuous everywhere. It should be noted that the bounding free streamlines in he case of the atmospheric jet are also lines of constant absolute vorticity.

6. Lee waves

It is clear that the problem considered by Long, i.e., the problem of the disturbance roduced by an obstacle in a stratified liquid, is very similar to the problem of atmospheric ee waves considered, for instance, by Scorer (1949). Scorer linearized his equations at the utset whereas Long studied certain velocity and density distributions that reduce the quation to a linear form, though amplitudes are unrestricted and the essential non-linear haracteristics of the solutions are preserved. If the Eqs. (9) are linearized we expect to et the Rossby wave equation (Rossby 1939) in the large-scale case and the lee-wave quation in the gravity-wave case, showing the direct analogy between the two types f wave. This is indeed the case as is indicated below.

Let U be the undisturbed velocity in the x direction and S the logarithm of the ndisturbed density, U and S being functions of y only. The linearized perturbation quations are :

$$U\nabla^2 v + U\frac{dS}{dy}\frac{\partial v}{\partial y} - \left(\frac{d^2U}{dy^2} - \beta + \frac{dU}{dy}\frac{dS}{dy}\right)v = 0 \quad . \quad . \quad (16a)$$

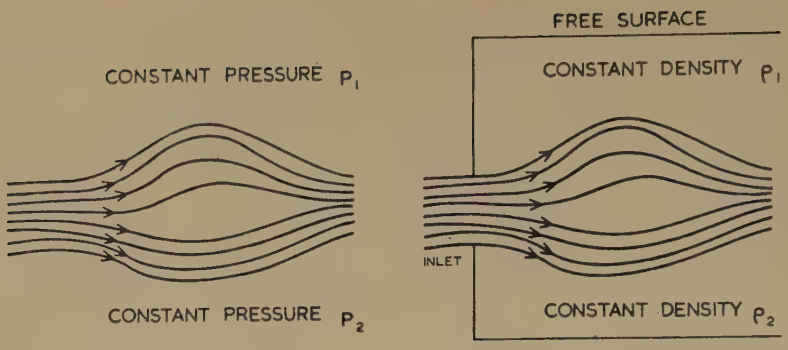
$$U\nabla^2 v + U\frac{dS}{dy}\frac{\partial v}{\partial y} - \left(\frac{d^2U}{dy^2} + \frac{g}{U}\frac{dS}{dy} + \frac{dU}{dy}\frac{dS}{dy}\right)v = 0 \quad . \quad . \quad (16b)$$

q. (16a) reduces to the familiar Rossby wave equation when the terms in dS/dy and $^2U/dy^2$ are neglected. This approximation is justified for the terms involving S, e.g. $\frac{U}{dy}\frac{dS}{dy}$ rarely exceeds $\beta/5$; on the other hand in many situations d^2U/dy^2 is *not* negligible ompared with β.

Perturbation equations for a stratified liquid similar to Eq. (16b) have been derived y Taylor (1931), Goldstein (1931) and Scorer (1951). Scorer (1949) also considered the erturbation equations for a *compressible* stratified fluid in connection with the study of ee waves. He derived an equation of the form :

$$U\nabla^2 v + U\left(\frac{dS}{dy} - \frac{g}{c^2}\right)\frac{\partial v}{\partial y} - \left(\frac{g}{U}\frac{dS}{dy} + \frac{d^2U}{dy^2}\right)v = 0 \quad . \quad . \quad (17)$$

here S is now the *potential* density. Note the appearance in this equation of the ' compressibility ' term g/c^2 in the coefficient of $\frac{\partial v}{\partial y}$, and the absence of the term $\frac{dU}{dy}\frac{dS}{dy}$ in the oefficient of v. This last term appears earlier in Scorer's analysis but is legitimately eglected for lee-wave applications since $g/U \gg dU/dy$. The solutions of Eq. (17) for arious boundary conditions have been discussed frequently in the last decade, perhaps he most recent contribution being that of Corby and Sawyer (1958).



(a) Atmospheric flow (b) Water-channel flow

Figure 3. Analogous flow patterns with free streamlines.

7. Conclusions

It has been pointed out that when a stably stratified liquid in horizontal motion is disturbed 'generalized vorticity' is conserved and the momentum tends to concentrate in meandering jets. The number of jets is determined by the Froude number, a quantity that is independent of the way in which the liquid is disturbed. The intensity of the jets and the amplitude of their meanders will however depend on the nature of the disturbance.

Similarly conservation of absolute vorticity in large-scale motions of the ocean and atmosphere can explain many observed features of the flow. In particular a number of the characteristics of 500 mb contour charts can be explained in this way, e.g., one or more jets with splitting and recombination, meandering motions and closed centres of circulation on both sides of the jet. The type of flow is determined by the dimensionless number $\beta L^2/U$, whereas the magnitude of the disturbance affects the amplitude of the oscillations and the intensity of the jets.

This inherent tendency of the fluid motion to organize itself into jets when the fluid is either statically stable because of its density distributions or dynamically stable because of its rotation, is apparently sufficiently marked to overcome the effect of turbulent mixing and friction. In the absence of some form of stability there is no tendency for jets to form and a jet which is injected into the flow dissipates rapidly.

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The introduction of the effects of topography into methods of numerical forecasting

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1. Introduction

The experience of weather forecasters makes it clear that mountain barriers have important effects on the movement and development of depressions and anticyclones. The distribution of land and sea also appears to modify the tracks of depressions. It is important, therefore, that the effects of topographic features should be incorporated in the models used as a basis for numerical weather prediction so far as this is possible. However, the dynamical mechanisms by which hills and mountains influence the large-scale weather systems are not yet fully understood, and it is the purpose of the present paper to survey the mechanisms most probably involved, to assess their relative importance and to indicate the manner in which appropriate modifications might be made to the quasi-geostrophic models of numerical forecasting.

The motions which are set up in an airstream as it crosses a mountainous area are not solely, or even primarily, on the horizontal scale which is treated by the equations used in numerical weather prediction, but it cannot be assumed that the smaller-scale motions have no effect on the behaviour of the large-scale systems for which predictions are to be made by numerical forecasting. The motions which are caused directly by topography can be classified broadly as follows :

(a) Small-scale turbulent motion mainly in the lower troposphere.
(b) Gravity oscillations which may extend upwards as far as the stratosphere.
(c) Oscillations about geostrophic equilibrium.
(d) Quasi-geostrophic motions.

The smaller-scale motions (a) and (b) dissipate energy and act as a resistance to the larger-scale motion. Only the quasi-geostrophic motions (d) can be adequately treated by the equations which currently form the basis of numerical weather prediction and even here there is a serious difficulty because the two-parameter model atmosphere, which is a convenient basis for computation, assumes a vertical distribution of the amplitude of the disturbance which is appropriate to depressions and anticyclones, but which may not necessarily be appropriate to orographically-induced disturbances.

The four scales of motion will be considered in the sections which follow.

2. Small-scale turbulent motion

It is convenient to think of the atmosphere as comprising a ' friction layer ' near the ground, some 400 to 800 m in depth, and the ' free atmosphere ' above. This is a well-defined meteorological concept and we may regard the friction layer as the part of the atmosphere in which the eddy-viscosity terms are important in the equations of motion and the free atmosphere as the part of the atmosphere in which they are negligible. The equations of numerical weather prediction which ignore the frictional terms are applicable only to the free atmosphere and numerical forecasting models should be designed to represent only the atmosphere above the friction layer. Moreover, since the eddy stress is to be regarded as negligible in the free atmosphere, it would be inappropriate to assume that the friction layer exerts a stress on the bottom of the free atmosphere, because the dynamical model provides no means of communicating this stress upward to any higher level.

How then does surface drag modify the large-scale motions of the free atmosphere? This question is discussed by Charney and Eliassen (1949) and their solution is both convenient for introduction into numerical forecasting, and probably describes the principal dynamical mechanism concerned. Charney argues as follows. Motion within the friction layer occurs with an approximate balance between the horizontal pressure gradient, the Coriolis acceleration, and the eddy stress. The result is the well-known drift across the isobars towards low pressure. There is necessarily a horizontal convergence of air within the friction layer into regions of cyclonic vorticity and divergence from regions of anticyclonic vorticity. A compensating net divergence or convergence must occur in the free atmosphere above the regions of cyclonic and anticyclonic vorticity in the lower layers. The positive correlation which thus results between divergence and vorticity in the free atmosphere leads to a gradual slowing down of the winds.

If this is the mechanism by which small-scale turbulence in the friction layer influences large-scale motion, its effects can be incorporated in numerical prediction methods by making allowance for vertical motion of the air through the lower boundary of the numerical forecasting model (now the top of the friction layer), and estimating this vertical velocity from the drift across the isobars in the friction layer. The problem becomes one of estimating the appropriate vertical velocities.

The suggested procedure is necessarily an imperfect representation of the real effects because there are certainly occasions, such as those of deep convection, when the eddy stress is not confined to a relatively shallow layer of less than 800 m. Also Sheppard, Charnock and Francis (1952) have shown that the eddy stress in the free atmosphere over the Atlantic is comparable with the surface stress. However, for numerical forecasting over periods of a day or two, surface friction over the oceans is probably too small to be important, and over the land where the surface stress is larger there is good evidence from the observed wind profiles that the eddy stress is considerably greater in the friction layer than above.

Estimates of the vertical velocity of the air at the top of the friction layer may be obtained from the equations of steady uniform motion, in which there is a balance between the pressure gradient force, the eddy stress and the Coriolis acceleration. For these conditions the equations of motion become

$$\left.\begin{aligned} f(u - u_g) &= \frac{1}{\rho}\frac{\partial \tau_y}{\partial z} \\ f(v - v_g) &= -\frac{1}{\rho}\frac{\partial \tau_x}{\partial z} \end{aligned}\right\} \quad . \quad . \quad . \quad (1)$$

where τ_x and τ_y are the horizontal components of eddy stress, u and v are the component velocities, and u_g and v_g are the corresponding geostrophic winds.

The geostrophic component of the flow contributes to the horizontal divergence and convergence only on account of the latitudinal variation of the Coriolis parameter. The corresponding divergence is $-\beta v_{gN}/f$ where β is the latitudinal variation of the Coriolis parameter f and v_{gN} is the northward component of geostrophic wind velocity.

The non-geostrophic part of the mass transport in the friction layer is obtained by integrating Eq. (1) from the ground, $z = 0$, to the top of the friction layer, $z = z_1$. Thus

$$\left.\begin{aligned} D_x &= \int_0^{z_1} \rho\,(u - u_g)\,dz = -\tau_{y0}/f = -\tau_0 \sin\alpha/f \\ D_y &= \int_0^{z_1} \rho\,(v - v_g)\,dz = \tau_{x0}/f = \tau_0 \cos\alpha/f \end{aligned}\right\} \quad . \quad . \quad (2)$$

Here it has been assumed that the eddy stress vanishes at the top of the friction layer and the x-axis has been taken along an isobar. The stress at the surface, τ_0, is assumed to be along the direction of the surface wind making an angle α with the isobars.

The value of ϖ_0 ($\equiv dp/dt$) corresponding to the vertical velocity at the top of the friction layer is obtained from the relation.

$$\varpi_0 = g \times \text{(total mass divergence in the friction layer)}$$

$$= g\left\{\frac{\partial D_x}{\partial z} + \frac{\partial D_y}{\partial y}\right\} - g\int_0^{z_1} \frac{\rho\beta v_{gN}}{f}\, dz$$

Assuming v_{gN} is approximately constant through the friction layer

$$\varpi_0 = \frac{g}{f}\left[\left\{\frac{\partial}{\partial y}(\tau_0 \cos\alpha) - \frac{\partial}{\partial x}(\tau_0 \sin\alpha)\right\} - \frac{\beta P}{g} v_{gN}\right] \quad . \quad . \quad (3)$$

where P is the pressure depth of the friction layer. The final term is usually smaller than the others.

Eqs. (2) have been used by several investigators, notably by Sutcliffe (1936) and Charnock, Francis and Sheppard (1956), to estimate the surface stress from the ageostrophic motion in the friction layer. Such estimates of τ_0 are particularly appropriate to the present problem but they have usually been expressed in the form

$$\tau_0 = k\,\rho\, V^2 \quad . \quad . \quad . \quad . \quad (4)$$

where V is the wind at anemometer level, and k is a non-dimensional constant determined by the investigation from the observed wind profiles. The surface drag determined in this way is considerably greater than that obtained by direct measurement of the force on the ground (e.g., Rider 1951) presumably because of the drag from larger obstructions such as buildings, bushes and trees.

For the purpose of incorporating the surface drag in numerical forecasting we need to relate τ_0 to the geostrophic wind velocity V_g rather than to the wind at anemometer level, and Sutcliffe's relation (4) has been transformed to

$$\tau_0 = k^* V_g^2 \quad . \quad . \quad . \quad . \quad (5)$$

by assuming $V = 0{\cdot}30\, V_g$, over land and $V = 0{\cdot}60\, V_g$ over the sea, together with an appropriate value of ρ. These relations between V and V_g have been based upon the comprehensive data given by Marshall (1954) with some corrections for the heights of the anemometers concerned. Figures given by Gordon (1950) suggest a slightly higher value of 0·71 for the ratio V/V_g over the sea.

Various representative wind profiles are given in the literature and values of τ_0 may be determined from them following the technique used by Sutcliffe. The corresponding estimates of k^* in Eq. (5) are given in Table 1 for comparison with the values deduced from the investigations of Sutcliffe and Charnock, *et al.*

TABLE 1. ESTIMATES OF k^* IN EQ. (5) (C.G.S. UNITS)

Author and reference	k^*	Location, etc. of observations
Land		
Sutcliffe (1936)	$0{\cdot}7 \times 10^{-6}$	England
Seeliger (1938)	$3{\cdot}3 \times 10^{-6}$	Northern Germany
Lettau (1950)	$1{\cdot}7 \times 10^{-6}$	Leipzig
Arkhangelskiy (1955)	$5{\cdot}0 \times 10^{-6}$	Moderately mountainous country in East Siberia
Buajitti and Blackadar (1957)	$5{\cdot}5 \times 10^{-6}$	Summer conditions in central U.S.A.
Sea		
Sutcliffe (1936)	$1{\cdot}8 \times 10^{-7}$	Atlantic
Charnock *et al.* (1956)	$0{\cdot}8 \times 10^{-7}$	Caribbean
Sverdrup (1957)	$18{\cdot}6 \times 10^{-7}$	Arctic ice

In calculating k^* from the data of Charnock *et al.*, the geostrophic wind has been taken as the mean in the lowest 300 m

The value of the frictional drag of the surface on the atmosphere depends upon the roughness of the surface. Apart from the large difference between the drag over the land and that over the sea, it may be necessary to take into account the greater frictional stress over rough and mountainous country when compared with more nearly level areas. The available estimates of the frictional drag based upon observations are inadequate to establish the magnitude of the effect, although Seeliger (1938) gives the change of wind direction with height in the friction layer for German stations divided according to their distances from the coast. His results suggest that the drag is greater over the rougher topography of central and southern Germany than over the northern plains in the ratio of 2 or 3 to 1.

Theoretical studies of the dependence of surface drag on surface roughness have been made by Rossby and Montgomery (1935), by Prandtl (1952) and by Ellison (1956). Despite widely different assumptions regarding the variation of the eddy diffusivity with height the authors agree that the surface drag increases only slowly with increase in surface roughness. The values of k^* (in Eq. (5)) have been estimated from the results given by the three authors for various roughness lengths and are given in Table 2 together with the corresponding deviation of the surface wind from the isobars α.

TABLE 2. THEORETICAL VARIATION OF FRICTIONAL DRAG WITH SURFACE ROUGHNESS

Author		Rossby and Montgomery		Prandtl		Ellison	
Terrain	z_0	k^*	α	k^*	α	k^*	α
Open grass land	3 cm	$1{\cdot}2 \times 10^{-6}$	19°	$2{\cdot}3 \times 10^{-6}$	27°	$2{\cdot}0 \times 10^{-6}$	9°
Agricultural land	10 cm	$1{\cdot}5 \times 10^{-6}$	21°	$3{\cdot}0 \times 10^{-6}$	29°	$2{\cdot}6 \times 10^{-6}$	10°
Rough terrain	1 m	$2{\cdot}2 \times 10^{-6}$	26°	$5{\cdot}0 \times 10^{-6}$	33°	$4{\cdot}3 \times 10^{-6}$	13°
Rugged terrain	10 m	$3{\cdot}6 \times 10^{-6}$	34°	$9{\cdot}0 \times 10^{-6}$	38°	$7{\cdot}5 \times 10^{-6}$	18°

The various estimates in Tables 1 and 2 differ considerably but a reasonable choice of trial values for k^* and α for use in experiments in numerical weather prediction would appear to be :

Over sea	$k^* = 2 \times 10^{-7}$	$\alpha = 6°$
Over land generally	$k^* = 1{\cdot}5 \times 10^{-6}$	$\alpha = 30°$
Over mountainous country	$k^* = 3 \times 10^{-6}$	$\alpha = 35°$

Values of α over sea and level land are taken from the studies of Marshall (1954) and the value for mountainous country has been assumed to be slightly greater.

In considering the effect on large-scale synoptic disturbances, it is desirable to consider how long the motion takes to adjust itself to a change in surface roughness. If the inertia terms are retained, Eqs. (1) become

$$\left.\begin{aligned} f(u - u_g) - \frac{Dv}{Dt} &= \frac{1}{\rho}\frac{\partial \tau_y}{dz} \\ f(v - v_g) + \frac{Du}{Dt} &= -\frac{1}{\rho}\frac{\partial \tau_x}{\partial z} \end{aligned}\right\} \quad . \quad . \quad . \quad (6)$$

Through the friction layer the terms in the right-hand side of Eq. (6) are of the order of $fV_g/4$ where V_g is the geostrophic wind velocity. If the first and last terms are out of balance because of a change in surface roughness the acceleration terms DV/Dt must be of the order of $fV/4$ and the time to bring about a change of velocity of the order $V_g/4$ and to approach frictional balance must be of the order of $1/f$, say 3 hr. This implies that with winds of 50 km hr^{-1} changes in surface roughness cannot be expected to have their

full effect until the air has travelled some 150 km across the new surface. Areas of increased roughness of smaller dimensions will be largely ineffective. Moreover, to affect the large-scale quasi-geostrophic disturbances the convergence within the friction layer must be spread over an area with dimensions which form a substantial fraction of the wavelength of a quasi-geostrophic disturbance. We are concerned thus with convergence which extends over areas with dimensions of 500 km or more, or which remains large enough to be effective when averaged over such an area.

To assess the magnitude of the vertical velocities arising from variations of surface roughness we may consider an air stream with velocity 20 m sec^{-1} flowing parallel to a mountainous coast. Over the sea we may regard the mass transport across the isobars in the friction layer as negligible, but over the land it is $\tau_0 \cos \alpha/f$ from Eq. (2), about 9 g cm^{-1} sec^{-1}. The value of ϖ_0 is obtained from Eq. (3) but must be smoothed over a distance of the order of 500 km. The value so obtained is 1·6 in c.g.s. units, equivalent to an upward velocity of 1·3 cm sec^{-1}. Such vertical velocities are not negligible when compared with the maximum vertical velocities which are computed from the quasi-geostrophic two-parameter models used in numerical weather prediction. These are of the order of 5 cm sec^{-1}.

An alternative measure of the effect of the frictional convergence upon the quasi-geostrophic motions is the vorticity change which may be expected to result, and this will be used to compare the frictional and other effects. The frictional convergence estimated in the previous paragraph corresponds to a rate of vorticity change averaged through the troposphere of $0{\cdot}22 \times 10^{-9}$ sec^{-2} (or $2{\cdot}8 \times 10^{-3}$ hr^{-2}).

3. Gravity oscillations

When an airstream crosses hilly country, vertical displacements of the current are frequently observed up to heights of several kilometres and even into the stratosphere; trains of leewaves sometimes form and the flow remains steady and non-turbulent. The lower streamlines roughly follow the ground profile and turbulent eddies are usually confined to the friction layer. The effect of the small-scale eddy motion on synoptic-scale disturbances has already been considered in the preceding section; we now need to consider the effects of the steady component of the flow.

In this section we consider only disturbances which are small enough for the effect of the earth's rotation to be neglected. This is possible if the horizontal dimensions of the hills and of the resulting disturbance are less than about 30 km. This may be seen as follows. The equations of motion for a steady uniform airstream crossing a ridge (velocity U in the undisturbed stream) lead to perturbation equations of the form :

$$\left.\begin{aligned} U\frac{\partial u'}{\partial x} - f v' &= -\frac{1}{\rho}\frac{\partial p'}{\partial x} \\ U\frac{\partial v'}{\partial x} + f u' &= 0 \end{aligned}\right\} \quad . \quad . \quad . \quad (7)$$

where u', v' and p' represent departures of the velocity components and pressure from their values in the undisturbed stream. Differentiation of the first of these equations and substitution in the second leads to

$$U\frac{\partial^2 u'}{\partial x^2} + \frac{f^2}{U}u' = \frac{1}{\rho}\frac{\partial^2 \rho'}{\partial x^2} \quad . \quad . \quad . \quad . \quad (8)$$

If the disturbance has a characteristic length L, $\partial^2 u'/\partial x'$ is of the order of u'/L^2 and the second term in Eq. (8) which represents the effect of the earth's rotation is small compared with the first if

$$L^2 \ll U^2/f^2 < U^2/10\,f^2$$

With $U = 10$ m sec^{-1} and $f = 1{\cdot}1 \times 10^{-4}$ sec^{-1} this requires $L < 30$ km.

The only comprehensive treatments of the steady 'gravity-type' disturbances of flow over hills are based on linear perturbation theory, but they give some useful insight into the nature of the flow. The streamlines are believed to have the general character illustrated in Fig. 1. The crests of the streamlines near the ridge occur along a line which is tilted from the vertical. The disturbance of the streamlines decreases in amplitude slowly with height. Leewaves may or may not be present on the downstream side of the obstacle.

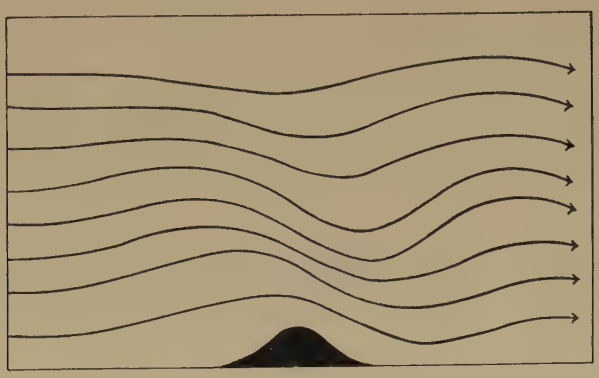

Figure 1. General character of streamlines over a ridge.

Some uncertainty attaches to the tilt of the streamline crests because, in almost all methods of calculating the flow so far adopted, the calculated tilt depends upon the mathematical boundary condition applied at the upper boundary of the area studied. Corby and Sawyer (1958) among others have argued that the appropriate boundary condition is one which leads in most natural airstreams to the displacement of the streamline crests upwind aloft as in Fig. 1. It is an important consequence of such a tilt that the downward-moving air has a greater horizontal velocity than the upward-moving air and that there is a downward transport of momentum towards the ground where a thrust is exerted upon the ridge by the higher pressure on the windward slope compared with the leeward slope. As we proceed upward, the downward transport of momentum must become smaller because the amplitude of the disturbance itself becomes smaller, and consequently there must be a loss of momentum from the airstream as it crosses the hill. If only the dynamics of large-scale systems are being considered this loss of momentum from the airstream may be replaced by a force acting within the body of the fluid to retard the motion.

Perturbation theory deals only with small departures of velocity from an 'undisturbed' velocity distribution, and it does not recognize any permanent change in the velocity of the airstream as a result of crossing a range of hills. The vertical transport of momentum through any level depends on the products of the perturbation velocity components, u' and v', such products being ignored in the dynamical equations as being of the second order of small quantities. Nevertheless, if a streamline field is calculated by perturbation methods it will give a valid estimate of the downward flux of momentum and its divergence. Complete solutions of the perturbation equations are not available in analytical form and numerical solutions appropriate to specific wind and mountain profiles are laborious to obtain. For the present purpose it will suffice to estimate the total drag upon the mountain or ridge arising from the gravity-type disturbance and to make a rough estimate of the depth of atmosphere which is losing momentum to the ground and throughout which the mountain is exerting a retarding influence.

An estimate of the drag upon a ridge may be obtained by considering a simple case of an airstream crossing a two-dimensional ridge. Scorer's parameter $l = \left(\frac{g\,\partial\theta/\partial z}{U^2\theta}\right)^{\frac{1}{2}}$ is considered independent of height, z. Following the treatment of Scorer (1949), the stream function is analysed into its Fourier components, such as $\zeta_k e^{ikx}$ and then the coefficient ζ_k of the component of wavelength $2\pi/k$ satisfies the equation

$$\frac{\partial^2 \zeta_k}{\partial z^2} + (l^2 - k^2)\,\zeta_k = 0 \qquad . \qquad . \qquad . \qquad . \qquad . \qquad (9)$$

The solution of Eq. (9) which satisfies the upper-boundary condition recommended by Corby and Sawyer (1958) is

$$\zeta_k = A\,e^{i(kx+\nu z)} \quad \text{where} \quad \nu^2 = l^2 - k^2 \qquad . \qquad . \qquad . \qquad (10)$$

The flow is here assumed to be in the positive direction of x.

If the profile of the ridge is represented by

$$H = \text{Real part } hb \int_0^\infty e^{ikx-kb}\,dk = \frac{hb^2}{x^2 + b^2}$$

the perturbation stream function ψ_0 at $z = 0$ must be

$$\psi_0 = \text{Real part } U_0\,hb \int_0^\infty e^{ikx-kb}\,dk \qquad . \qquad . \qquad . \qquad (11)$$

because $\qquad \partial\psi_0/\partial x = w_0 = U_0\,\partial H/\partial x$

Using Eq. (11) as the lower-boundary condition, A in Eq. (10) is determined, and the stream function ψ_z at level z is given by

$$\begin{aligned}\psi_z &= \int_0^\infty \zeta_k\,dk \\ &= U_0\,hb \left\{ \int_0^l e^{-bk} \cos\{kx + \nu z\}\,dk + \int_l^\infty e^{-bk-(k^2-l^2)^{\frac{1}{2}}z} \cos kx\,dk \right\}\end{aligned}$$

At $z = 0$ the perturbation velocities u_0 and w_0 are given by

$$u_0 = -\left(\frac{\partial\psi_z}{\partial z}\right)_0 = U_0\,hb \left\{ \int_0^l \nu e^{-bk} \sin kx\,dk + \int_l^\infty (k^2 - e^2)^{\frac{1}{2}} e^{-bk} \cos kx\,dk \right\}$$

$$w_0 = U_0 \frac{\partial H}{\partial x} = U_0\,hb \frac{d}{dx}\left(\frac{b}{b^2 + x^2}\right)$$

The total drag of the air on the mountain is obtained by integrating the momentum transport $-\rho\,u_0\,w_0$ along the x-axis.

$$\begin{aligned}\text{Drag} &= -\rho \int_{-\infty}^\infty u_0\,w_0\,dx \\ &= -\rho \int_{-\infty}^\infty dx \left[U_0^2\,h^2\,b^2 \frac{d}{dx}\left(\frac{b}{b^2 + x^2}\right) \left\{ \int_0^l e^{-bk} (l^2 - k^2)^{\frac{1}{2}} \sin kx\,dk \right.\right. \\ &\qquad \left.\left. + \int_l^\infty e^{-bk} (k^2 - l^2)^{\frac{1}{2}} \cos kx\,dk \right\} \right] \\ &= -\rho\,U_0^2\,h^2\,b^2 \left\{ \int_0^l (l^2 - k^2)^{\frac{1}{2}} e^{-bk} \int_{-\infty}^\infty \sin kx \frac{d}{dx}\left(\frac{b}{b^2 + x^2}\right) dx\,dk \right. \\ &\qquad \left. + \int_l^\infty (k^2 - l^2)^{\frac{1}{2}} e^{-bk} \int_{-\infty}^\infty \cos kx \frac{d}{dx}\left(\frac{b}{b^2 + x^2}\right) dx\,dk \right\}\end{aligned}$$

The second integral vanishes and the integral with respect to x in the first expression can be evaluated by integration by parts followed by contour integration. The following expression for the drag is thus obtained

$$\text{Drag} = \pi\rho\, U_0^2\, h^2\, b^2 \int_0^l k\,(l^2 - k^2)^{\frac{1}{2}}\, e^{-2bk}\, dk \quad . \quad . \quad . \quad (12)$$

Numerical evaluation of the integral (12) shows that for $b = 1/l$ the drag is $0{\cdot}11\ \pi\rho\ U_0^2\, h^2\, l$ and that with increasing width of the ridge the drag increases up to $0{\cdot}25\ \pi\rho\ U^2\, h^2\, l$.

In order to estimate the drag on a typical airstream we may put $h = 300$ m, $U_0 = 20$ m sec^{-1}, $l = 0{\cdot}5$ km^{-1}, $b = 2$ km and $\rho = 1{\cdot}23 \times 10^{-3}$ g cm^{-3}. The resulting drag is $7{\cdot}6 \times 10^{-6}$ dyne cm^{-1}. If the ridges are spaced at intervals of 10 km the average drag is 7·6 dyne cm^{-2} which is about the same as the frictional drag over ordinary terrain in a similar wind.

In order to assess from what levels in the atmosphere momentum is being lost, and upon which an effective drag is being exerted, it would be necessary to carry out a similar evaluation of the downward momentum transport at several levels in the atmosphere. This does not admit of so simple a solution as that obtained for ground-level, and since the value and distribution of the drag will vary considerably with the wind and temperature distribution in the airstream it will probably suffice to base an estimate of the vertical distribution of the drag on the manner in which the amplitude of the disturbance decreases with height, because the downward transport of momentum must in general decrease as the amplitude decreases.

The decrease in amplitude of the streamline disturbance with height can be looked upon as arising from two causes. The first is the finite width of the ridge which implies that when the profile is analysed into Fourier components each component wave train has a different tilt to the vertical. As we proceed upward the contributions of the different components differ in phase and interfere, so that the total amplitude decreases. The second factor leading to a decrease of amplitude with height is the limited length of real ridges transverse to the airstream.

The second of these two effects is illustrated in a discussion of the airflow over an isolated mountain by Scorer (1956). Scorer calculates the streamlines over both circular and elongated hills. Examination of his diagrams gives a rough indication of the level at which the amplitude is about half its value near the ground. For a circular hill of a few kilometers diameter this appears to occur at a level for which $lz = 2{\cdot}5$ and for an elongated hill where $lz = 4{\cdot}5$, l being Scorer's parameter which has typical values of the order of 0·5 or 1 km^{-1}. Recalling that the vertical transport of momentum will vary with the square of the amplitude of the disturbance and that there is an additional effect of the finite width of the mountain which does not appear in Scorer's results because of an approximation in the calculations and which tends to reduce the disturbance aloft, it seems reasonable to assume that roughly half of the effective drag is exerted on the lowest 3 km of the atmosphere and the remainder is distributed over the upper troposphere and stratosphere.

A rough approximation to the effect on the atmosphere of the drag from gravity-type disturbances would thus be provided by introducing into the equations of motion a force in the opposite direction to the low-level wind and decreasing linearly from a maximum near the ground to zero at the tropopause. This will necessarily be a crude approximation but the actual effect will vary so much with the character of the terrain, and the vertical variation of wind and temperature in the airstream, that any approximation which attempted a closer degree of realism would have to take into account details of the airstream which do not form part of the current models of numerical weather prediction. The appropriate values of the total drag might be estimated from the expression (12) using values of U_0 appropriate to the occasion, a value of h appropriate to the terrain and conventional values for b, ρ and l.

Introduction of the drag into a quasi-geostrophic model for numerical weather prediction would involve an extra term in the vorticity equation. For example, Eq. (6) of the treatment by Sawyer and Bushby (1953) would become

$$\frac{\partial}{\partial t}(\zeta_g + f) + V_g \operatorname{grad}(\zeta_g + f) + \varpi \frac{\partial}{\partial p}(\zeta_g + f) + \frac{1-\alpha}{p_0 - p_1} g \operatorname{curl}_s F = (\zeta_g + f)\frac{\partial \varpi}{\partial p}$$

where F is the total drag exerted by the airstream upon the terrain through the medium of gravity-type disturbances. The effect of the introduction of this new term in the vorticity equation is to insert additional terms into Eqs. (26) and (27) of Sawyer and Bushby's paper as follows :

(a) on the left-hand side of the equation for $\partial h_m/\partial t$ namely Eq. (27) :

$$-f \operatorname{curl}_s F/(p_0 - p_1)$$

(b) on the left-hand side of the equation for $\partial h'/\partial t$ namely Eq. (26) :

$$f \operatorname{curl}_s F/(p_0 - p_1)$$

In order to assess $\operatorname{curl}_s F$ we may imagine a mountainous region exerting a drag of 7·6 dyne cm^{-2} (as estimated above) and an adjacent area which produces no topographic resistance. As in estimating the effect of surface friction, only effects over distances of the order of 500 km are material, so that the effective value of $\operatorname{curl}_s F$ is of the order of $1{\cdot}5 \times 10^{-7}$ g sec^{-2} cm^{-2}. The rate of change of vorticity arising from the term $f \operatorname{curl}_s F/(p_0 - p_1)$ is then

$$\begin{aligned} g \operatorname{curl}_s F/(p_0 - p_1) &= 1{\cdot}3 \times 10^{-10} \text{ sec}^{-2} \\ &= 2{\cdot}4 \times 10^{-3} \text{ hr}^{-2} \end{aligned}$$

This is very nearly the same as the vorticity change estimated as arising from surface friction effects in § 2.

4. OSCILLATIONS ABOUT GEOSTROPHIC EQUILIBRIUM

Quasi-horizontal oscillations of air about geostrophic equilibrium with a period of the order of 18 hr or so have been the subject of an extensive theoretical literature, but although such oscillations have been observed in the trajectories of constant-level balloons by Neiburger and Angell (1956), there is no evidence that they have any systematic effect on the behaviour of large-scale meteorological systems.

Queney (1948) has shown theoretically that such oscillations can be set up by a mountain ridge if its width is of the order of 100 km. He gives an illustration which shows the streamlines (computed for small amplitudes) over such a mountain. The mountain is associated with a ridge of high pressure in the sea-level isobars, and the airstream is deflected to the left over the windward slope and to the right over the leeward slope. Superimposed on the motion are a series of oscillations about geostrophic equilibrium with a wavelength of the order of a few hundred kilometers. The vertical displacements of the streamlines as calculated by Queney appear to be similar to character in those calculated for gravity oscillations over a mountain of smaller width, and there does not seem any reason to suppose that the downward flux of momentum and resulting drag will be of a different order of magnitude from that calculated from Eq. (12). However, in a given space we can have only one ridge of 100 km width for ten of 10 km width and it seems likely that the effective drag resulting from these longer wavelength gravity-Coriolis oscillations will be considerably smaller than the drag arising from the shorter wavelength oscillations of a simple gravity type.

Since ageostrophic motion is ignored in the prediction equations used in quasi-geostrophic formulations of the prediction equations of numerical forecasting, it is impossible to introduce into them any effect of the horizontal non-geostrophic motions set up by mountains except possibly as an additional stress arising from a type of horizontal eddy transfer of momentum. No estimate of the magnitude of these ageostrophic winds is

available, but it seems unlikely that they would exceed 5-10 kt except within 100 km or so of the mountain slopes. The resulting horizontal eddy stress is likely to be of minor importance.

The overall displacement of the air to the left as it crosses the ridge and its subsequent return to the right which are shown in Queney's diagrams are essentially quasi-geostrophic phenomena, and this aspect of the motion will be largely represented by the quasi-geostrophic treatment of the largest-scale disturbances which are discussed in § 5 which follows.

5. Quasi-geostrophic motions

(a) *Small amplitudes*

If a mountain ridge is sufficiently broad (500-1,000 km in width) or if the smoothed ground profile has components with appropriately long wavelengths, an airstream as it crosses the ridge will have areas of vertical motion of sufficient extent to be of direct importance in the quasi-geostrophic equations of numerical weather prediction. This effect of the mountain ridge can be introduced into the prediction equations by a simple modification of the lower-boundary condition, and this is what has been done in nearly all attempts made so far to introduce topographic effects into numerical weather prediction. Examples of treatment of the orographic effect in this way are found in papers by Estoque (1957) and by Brandejs and Vitek (1957).

However, as most numerical forecasts are based upon an incomplete vertical representation of the vertical variation of the variables, it becomes necessary to specify how the vertical velocities set up in the air as it crosses a mountain ridge themselves vary with height. Some insight into this problem can be obtained by considering a very simple case. Suppose that a disturbance of small amplitude is set up by a broad mountain ridge extending across a steady uniform airstream with velocity U.

The problem is made two-dimensional by assuming that there are no variations in the y-direction parallel to the mountain ridge. The horizontal coordinate is then x and pressure is used as a vertical coordination, p.

The continuity and vorticity equations simplify and combine as follows :

$$\frac{\partial \varpi}{\partial p} = -\operatorname{div}_z V = -\frac{1}{(\zeta+f)}\frac{D\zeta}{Dt} = -\frac{1}{(\zeta+f)}\frac{D}{Dt}\left(\frac{\partial v}{\partial x}\right)$$

$$= -\frac{1}{(\zeta+f)}\left\{U\frac{\partial}{\partial x}\left(\frac{\partial v}{\partial x}\right) + \varpi\frac{\partial}{\partial p}\left(\frac{\partial v}{\partial x}\right)\right\} \quad . \quad . \quad . \quad (13)$$

where v is the velocity parallel to the ridge, ζ is the vertical component of vorticity and $\varpi \equiv Dp/Dt$.

Omitting all terms of the second order in the perturbation quantities Eq. (13) reduces to

$$\frac{\partial \varpi}{\partial p} = -\frac{U}{f}\frac{\partial^2 v}{\partial x^2} \quad . \quad . \quad . \quad . \quad (14)$$

The 'thermal' wind relation is obtained by differentiating the geostrophic wind relation with respect to p.

Thus

$$\frac{\partial v}{\partial p} = \frac{g}{f}\frac{\partial}{\partial p}\left(\frac{\partial h}{\partial x}\right)$$

$$= -\frac{1}{f}\frac{\partial}{\partial x}\left(\frac{1}{p}\right)$$

$$= -\frac{R\,p^{-\frac{1}{\gamma}}}{f\,p_0^{\frac{\gamma-1}{\gamma}}}\frac{\partial \theta}{\partial x} \quad . \quad . \quad . \quad . \quad (15)$$

where θ is potential temperature and p_0 is standard pressure. If the motion is adiabatic, to the order of approximation in use

$$\frac{D\theta}{Dt} = U\frac{\partial\theta}{\partial x} + \varpi\frac{\partial\theta}{\partial p} = 0 \qquad (16)$$

Combining (15) and (16)

$$\frac{\partial v}{\partial \rho} = \frac{R\,p^{-\frac{1}{\gamma}}}{f\,U\,p_0^{\frac{\gamma-1}{\gamma}}}\frac{\partial\theta}{\partial p}\varpi \qquad (17)$$

Differentiating (14) with respect to p and substituting from (17)

$$\frac{\partial^2\varpi}{\partial\rho^2} = -\frac{R\,\rho^{-\frac{1}{\gamma}}}{f^2\,p_0^{\frac{\gamma-1}{\gamma}}}\frac{\partial\theta}{\partial p}\frac{\partial^2\varpi}{\partial x^2} = B^2\frac{\partial^2\varpi}{\partial x^2} \qquad (18)$$

where $$B^2 = \frac{1}{g\,\rho\,f^2}\frac{1}{\theta}\frac{\partial\theta}{\partial z}$$

Approximately $\frac{\partial^2}{\partial p^2} = \frac{1}{g^2\rho^2}\frac{\partial^2}{\partial z^2}$ and for a harmonic disturbance of the form $\varpi = \varpi^* e^{ikx}$ Eq. (18) becomes

$$\frac{\partial^2\varpi^*}{\partial z^2} + B^2k^2g^2\rho^2\varpi^* = 0$$

It follows that a measure of the vertical scale of the disturbance is $2\pi/Bkg\rho$ (equal to the vertical wavelength if $B\rho$ is constant), and the vertical scale of the disturbance is of the order of $1/Bg\rho$ times the horizontal scale. For a representative value of the stability $\left(\frac{1}{\theta}\frac{\partial\theta}{\partial z} = 1{\cdot}8\times10^{-7}\,\mathrm{cm}^{-1}\right)$ this ratio becomes $8{\cdot}3\times10^{-3}$, and can be taken as a rough indication of the ratio of the vertical to the horizontal scale of the quasi-geostrophic disturbance set up by a mountain ridge. The vertical motion over a ridge with a width of 1,000 km will decrease upward and become small at about the top of the troposphere. The effects of narrower ridges will decrease more rapidly with height and of broader ridges more slowly, but the assumption that ϖ decreases linearly from the ground to the tropopause appears to be a reasonable representation of the effect of the major topographic features whose effects we wish to introduce into numerical forecasting methods in this way. This assumption has already been widely used.

Gambo (1957) has used a similar discussion to that of the preceding paragraphs to demonstrate that, if the convergence and divergence associated with orographic disturbances is to be sufficiently small for them to be treated as quasi-geostrophic, then the slope of the ground must not exceed 1/1,000.

A wind of 20 m sec^{-1} crossing a mountain ridge of width 1,000 km and height 300 m produces a vertical velocity of the order of 1 cm sec^{-1}. If the compensating divergence falls off upward to zero at the tropopause the mean divergence is about $3{\cdot}5\times10^{-6}$ sec^{-1} and the resulting vorticity change about $3{\cdot}5\times10^{-10}$ sec^{-2} (3×10^{-3} hr^{-2}). This is slightly greater than the vorticity changes estimated as due to the gravity-wave drag from a series of ridges of similar height, but since the simple lifting effect of the mountain produces equal and opposite vorticity changes in the same air parcel as it ascends and descends the two sides of the mountain, the large-scale effect may in fact be smaller than that of the 'gravity-wave' and frictional drags which produce opposite vorticities in different elements of the atmosphere.

(b) *High mountains*

Throughout the preceding discussion it has been assumed that the mountains were of a height small enough to permit all of the air to pass over them. Little progress has been made on the general hydrodynamical theory appropriate to motions of larger amplitude and it is difficult to provide a theoretical basis for considering the motion when a substantial part of the air is deflected round the sides of the mountain mass.

However, some numerical experiments might be conducted to study the extreme case in which the mountain range acts as a complete barrier to the air in the lowest layers of the atmosphere. If a two-parameter model is used as a basis for computation it would be appropriate to assume that the flow at 1,000 mb should skirt the mountain barrier. It is uncertain to what extent the 600 mb flow would be directly affected, but, as a trial hypothesis, it might be assumed that there is no vorticity change directly produced by vertical motions set up by the mountain.

The method of solution of the two-parameter equations would then follow the following lines. The prediction equations would be re-written as a Poisson equation for $\partial h_m/\partial t$ (h_m = 600 mb height) which would be solved first, together with a Helmholtz equation for $\partial h_0/\partial t$ (h_0 = 1,000 mb height). The solution for $\partial h_0/\partial t$ would then be obtained by the usual relaxation procedure subject to additional conditions which require that h_0 should be the same at all points of the boundary of the mountain mass and that the circulation around the mountain mass at level h_0 remains unaltered. It is uncertain whether a relaxation procedure based on these conditions will converge, but, expressed as a problem in linear equations, there appear to be the required number of equations for the unknowns.

If such a modification of numerical forecasting procedure gives results which reproduce significant features of atmospheric behaviour, a more realistic treatment may be possible by assuming that only part of the flow skirts the mountain and combining the treatments of this and the preceding sub-section.

6. Discussion and conclusions

It appears from the preceding discussion that topography is likely to affect the behaviour of large-scale weather systems through several different mechanisms. The additional terms in the prediction equations which are needed to represent such topographic effects are smaller than the principal terms in the equations by a factor of only 5 to 10 and may on occasion be sufficiently large to be of primary importance.

It is possible to distinguish at least three different mechanisms by which topography exerts a significant effect on the large-scale flow. These may be regarded as arising (a) from surface friction in the boundary layer, (b) from drag exerted by gravity-waves in the troposphere and (c) from quasi-geostrophic effects. The frictional effect (a) arises from features of the ground profile from a few centimetres to 1 km in extent; gravity-wave drag (b) arises from features of the topography with dimensions of a few kilometres, and the quasi-geostrophic effects (c) depend on variations of ground elevation extending over hundreds of kilometres. The relative magnitudes of these components in the ground profile vary greatly from place to place and the relative importance of the three effects will have corresponding variations. However, the estimates of magnitude which have been made suggest that none of the effects is always dominant. The frictional effect is likely to extend over wider areas than the others and the transition from land to sea may be of as great (or greater) importance in regard to surface friction as any changes in the roughness of the terrain.

A scale of orographic disturbance intermediate between (b) and (c) has been omitted from detailed consideration. This involves non-geostrophic components of the motion on a scale comparable with fronts and other small synoptic systems. It is not possible to consider the introduction of such topographic effects into numerical weather prediction until a satisfactory non-geostrophic model has been developed and tested as a basis for numerical integrations.

The elementary discussion of the present paper has not considered the nature of the changes in large-scale flow pattern which can be expected from the mechanisms of topographic influence which have been discussed. There is a need for this to be studied both by means of numerical integrations of the equations starting from initial conditions representing weather situations as they occur and also in respect of much simplified and idealised flow patterns and mountain profiles.

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The collision efficiency of small drops

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Summary

The forces on two moving spheres are found from the Stokes equations of motion, full allowance being made for the mutual interference of the flows round the spheres, and the relative trajectories when they are falling under gravity are calculated so that the collision efficiency E can be found. The values of E are determined for drops of radius up to 30 μ colliding with droplets of any smaller size. The results differ widely from those found by Pearcey and Hill (1957) and it is suggested that the reason for the discrepancy is that the motion of the spheres when they are close together is of great importance in determining E. The results also differ from those found by Schotland and Kaplin (1956). Their experiments did not model the ratio of the density of the spheres to that of the medium through which they are falling and since it is shown that this factor is of importance, their results are not considered applicable to water drops falling in air.

1. Introduction

In determining the theoretical rate-of-growth of drops by their coalescence with smaller droplets in a non-freezing cloud, it is necessary to know the collision efficiency, E. This quantity is defined for a drop and droplet of known radii as follows. If the drop and droplet are falling in isolation in still air, the centre of the droplet must lie within a certain horizontal circle a great distance below the drop if it is not to be swept round the drop but collide with it. The ratio of the area of this circle to the cross-section of the drop is E. It is convenient to denote the larger and smaller drops by the terms drop and droplet respectively even when they are nearly equal in size.

Various attempts have been made to calculate E, all of which use the method of determining the forces on the drop and droplet when they are moving with arbitrary velocities in any relative position by hydrodynamical theory, so that the equations of motion of the drop and droplet can be found. Starting with a large initial vertical separation of the drop and droplet, the equations of motion are solved to determine the relative trajectory. By varying the initial horizontal separation, the grazing trajectory with which the two spheres just collide can be found and the value of E determined. The difficulty in this calculation is that of determining the forces. This is simplified if the drop is very much larger than the droplet and a summary of the results found with this restriction will be found in Mason (1957). A recent paper by Pearcey and Hill (1957) is the first theoretical attempt to calculate E for collisions between drops of comparable size. They give values of E for drops in the range 4 μ to 200 μ radius. Their method is to start with the Oseen flow for a single moving sphere in the fluid at rest. The flow pattern for the two moving spheres is found simply by superposition of the two individual flow patterns for the motion of the spheres in isolation. If $\mathbf{U}$ is the velocity of one of the spheres and $\mathbf{u}$ the velocity produced at the point occupied by the centre of that sphere by the motion of the other sphere in isolation, the drag force is taken to be proportional to $\mathbf{U} - \mathbf{u}$, the constant of proportionality being determined by the experimental value of the drag force on a sphere of the same size moving with a uniform velocity $|\mathbf{U} - \mathbf{u}|$ in the fluid at rest. The approximations involved in this procedure are justifiable when the spheres are widely separated, but when the spheres are close together this simple method of allowing for the mutual influence of the spheres is considerably in error; also the Oseen approximation itself is poor in the neighbourhood of the drop surfaces giving an additional source of

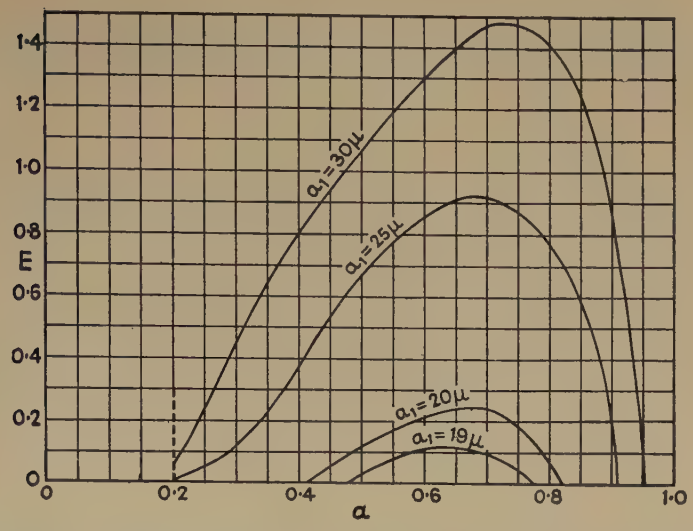

Figure 1. The collision efficiency of small drops. A drop of radius a_1 microns collides with droplets of radius a_2 microns lying in a horizontal circle of area E times the cross-section of the drop. The values of E are shown for various drop radii and for different $a = a_2/a_1$.

error in the region of near contact. Pearcey and Hill justify their neglect of these considerations by stating that the droplet's inertia will be sufficient to carry it rapidly through this region. It is shown below, however, that there are cases when the droplet is swept round the drop remaining close to it for some time (see Fig. 2).

The method of the present paper is to use the Stokes approximation to the hydrodynamical equations, the complete solution of which for the general motion of two spheres can be found. This restricts the size of the drops which can be considered to a radius of 30 μ or less but takes account of the mutual interference of the flows even when the spheres are close together. The results found are shown in Fig. 1.

2. The equations of motion

As the Stokes equations are linear, the flow for two spheres moving with arbitrary velocities can be found by superimposing the flows for two spheres moving along and perpendicularly to their line of centres. The method used to determine the drag forces in the two motions can be found elsewhere (Hocking 1958), so only the results will be given here. If the radii of the spheres are a_1 and a_2 and the distance between their centres is r, the drag forces can be found in powers of a_1/r and a_2/r to as high a degree as required. It was considered sufficient to take the the terms up to and including the seventh power, the next three terms making very little difference even when the spheres are close together. The accuracy decreases when the spheres are of very different sizes and close together, so that the ratio of the radii, a_2/a_1, for which the expressions for the drag forces can be used should be greater than about 0·2.

If the velocities of the spheres along the line of centres are U_1 and U_2 and η is the viscosity of air, the drag forces on the spheres are in the direction of the line of centres and to the accuracy required are $-6\pi\eta\, a_1\, L_1$ and $-6\pi\eta\, a_2\, L_2$, where

$$L_1 = L_1(U_1, U_2, a_1, a_2, r) =$$

$$= \frac{U_1(4r^7 - 15a_1^3a_2r^3 - 8a_1^5a_2r + 30a_1^3a_2^3r) - U_2(6a_2r^6 - 2a_2(a_1^2+a_2^2)r^4 + 75a_1^3a_2^4)}{4r^7 - 9a_1a_2r^5 - 9a_1a_2(a_1^2+a_2^2)r^3 - 9a_1a_2(a_1^4+a_2^4)r + 58a_1^3a_2^3r} \quad (1)$$

and $L_2 = L_1(U_2, U_1, a_2, a_1, r)$.

Similarly, if the velocities of the spheres are V_1 and V_2 perpendicular to their line of centres, the drag forces are perpendicular to the line of centres and to the same accuracy as above are $-6\pi\eta\, a_1\, T_1$ and $-6\pi\eta\, a_2\, T_2$, where

$$T_1 = T_1(V_1, V_2, a_1, a_2, r)$$

$$= \frac{V_1(16r^7 - 12a_1^3a_2r^3 - 17a_1^5a_2r) - V_2(12a_2r^6 + 4a_2(a_1^2 + a_2^2)r^4 + 6a_1^3a_2^4)}{16r^7 - 9a_1a_2r^5 - 18a_1a_2(a_1^2 + a_2^2)r^3 - 18a_1a_2(a_1^4 + a_2^4)r - 2a_1^3a_2^3r} \qquad (2)$$

and $T_2 = T_1(V_2, V_1, a_2, a_1, r)$.

Some idea of the behaviour of the functions L_1, and T_1, can be obtained if only the higher powers of r are retained. Thus, neglecting the terms in $1/r^3$, which is a valid approximation when the spheres are 5 or more radii apart, the values of L_1 and L_2 are

$$L_1 = U_1 - 3\,a_2\,U_2/2\,r + 9\,a_1\,a_2\,U_1/4\,r^2$$

$$L_2 = U_2 - 3\,a_1\,U_1/2\,r + 9\,a_1\,a_2\,U_2/4\,r^2.$$

The first terms are those that remain when the spheres do not influence each other, the second terms give the reduction in the drag forces on each sphere due to the motion of the other and the third terms give the increase in the drag forces due to the modification of the flow round each sphere caused by the presence of the other. The first of these modifications is greater for the smaller sphere than for the larger and for the slower sphere than for the faster, but the second and less important modification, which falls off like $1/r^2$, is proportionately the same for both spheres. The same two effects of the mutual influence of the spheres are found for the transverse motion but the corresponding terms in T_1 and T_2 are reduced by a factor of 2 for the first effect and 4 for the second.

Before finding the equations of motion it is convenient to make the variables non-dimensional. The unit of length is a_1, the radius of the larger sphere, and the radius of the smaller sphere is denoted by $a = a_2/a_1$. The unit of velocity, U, is the terminal velocity of the larger sphere falling by itself under gravity and is given by $U = 2\,\sigma\, g a_1^2/9\eta$, where σ is the density of the sphere and g is the gravitational acceleration. The hydrostatic upthrust can be ignored for water drops in air. The terminal velocity of the smaller sphere is Ua^2. Taking the x-axis vertically downwards and the y-axis horizontal, the coordinates of the centre of the sphere of radius 1 in the non-dimensional system is (x_1, y_1) and its velocity components are (u_1, v_1); the corresponding quantities for the sphere of radius a have a suffix 2. As only the relative trajectory is required, the difference of the coordinates of the centres is all that is needed, i.e., $x_2 - x_1 = x$ and $y_2 - y_1 = y$, which determine the vertical and horizontal separations of the spheres. The distance between the centres r, is given by $r^2 = x^2 + y^2$. The vertical and horizontal components of the drag forces can be found in terms of these velocity components and positions and, with the weight included, can be equated to the mass-accelerations of the spheres, giving

$$\left.\begin{aligned} I\frac{du_1}{dt} &= 1 - \frac{l_1 x - t_1 y}{r} \\ I\frac{du_2}{dt} &= 1 - \frac{l_2 x - t_2 y}{a^2 r} \\ I\frac{dv_1}{dt} &= -\frac{l_1 y + t_1 x}{r} \\ I\frac{dv_2}{dt} &= -\frac{l_2 y + t_2 x}{a^2 r} \end{aligned}\right\} \qquad (3)$$

and the relations between the relative position and the velocities are

$$\frac{dx}{dt} = u_2 - u_1$$

$$\frac{dy}{dt} = v_2 - v_1$$

In these equations, I is a dimensionless parameter defined by

$$I = \frac{4\,\sigma^2\, g a_1^3}{81\,\eta^2} \qquad . \qquad . \qquad . \qquad . \qquad . \qquad (4)$$

and, with their values given by Eqs. (1) and (2),

$$l_1 = L_1\,(\alpha_1, \alpha_2, 1, a, r)$$

$$l_2 = L_1\,(\alpha_2, \alpha_1, a, 1, r)$$

$$t_1 = T_1\,(\beta_1, \beta_2, 1, a, r)$$

$$t_2 = T_1\,(\beta_2, \beta_1, a, 1, r)$$

where

$$\alpha_1 = \frac{u_1 x + v_1 y}{r}, \qquad \alpha_2 = \frac{u_2 x + v_2 y}{r}$$

$$\beta_1 = \frac{v_1 x - u_1 y}{r}, \qquad \beta_2 = \frac{v_2 x - u_2 y}{r}$$

At 0°C, $I = 0{\cdot}00166\ a_1^3$, where a_1 is measured in microns. I is independent of pressure but not of temperature; if the results are to be applied at different temperatures, Eq. (4) will determine the drop radius which will leave the value of I unaltered at the new temperature.

It is supposed that initially the spheres are falling with their terminal velocities so that $u_1 = 1$, $u_2 = a^2$, $v_1 = v_2 = 0$. The initial vertical separation was taken as $x = 50$. To test whether this was sufficiently large, one trajectory was calculated with $x = 100$ initially and no significant difference was found. The initial value of the horizontal separation y was varied until the trajectory which made the smallest value of r, the distance between the centres, equal to $1 + a$, the sum of the radii, was found. If this critical value of y is denoted by y_c, collision occurs if $y \leqslant y_c$ initially and the collision efficiency, E, is y_c^2.

The method used for the solution of the equations of motions was a Runge-Kutta method of step-by-step integration and the calculations were performed on the DEUCE electronic computer at the London Computing Centre of the English Electric Company. Four trajectories were usually sufficient to enable E to be found to two decimal places, particularly after several results had been found. The time taken for the computation of a trajectory varied with the amount of information produced on punched cards by the machine. If only the minimum distance between the spheres was required, the time taken was about 8 min but if the shape of the trajectory when the spheres were close together was also required, the time was increased to about 13 min.

3. Discussion of the results

The values of E, shown in Fig. 1, were found for drop sizes up to 30 μ radius and for all droplet sizes provided that $a \geqslant 0{\cdot}2$, the reason for the exclusion of smaller values of a being explained above. The choice of 0·2 as the limiting value for a was suggested by calculations for smaller values of a and for drop sizes of 25 μ and 30 μ without including any extra terms in Eqs. (1) and (2). If the results found were included in Fig. 1, the curves for these drop sizes would flatten out in the region $a < 0{\cdot}2$ to give E a roughly constant value of 0·004. This predicted change in the shape of the curves is probably

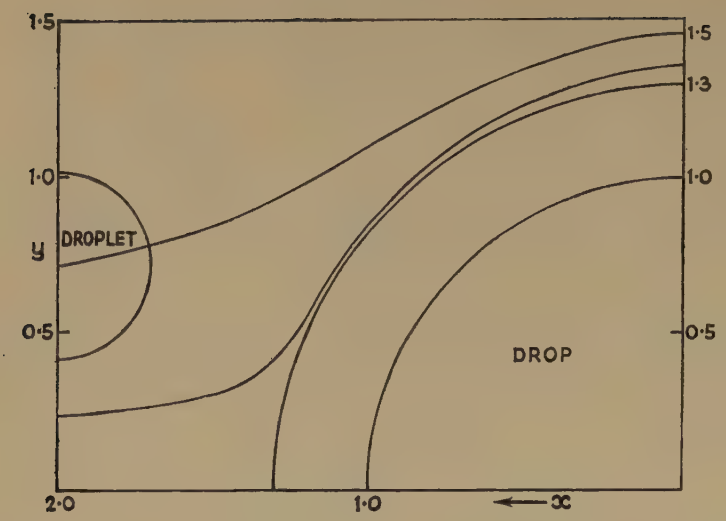

Figure 2. Two trajectories for a drop of radius 20 μ and a droplet of radius 6 μ. The initial values of y are 0·47 and 0·15. The trajectories are drawn relative to the drop and the sizes of the drop and droplet are shown, also the circle that the trajectory must touch if collision is to occur.

the result of the errors involved in using Eqs. (1) and (2) for very small values of a; in reality one would expect the 25 μ and 30 μ curves to show a value for a below which collision cannot occur as with the curves for other drop radii. However, since even the approximate calculations predict a value of E which is only 0·004, this point is not of great practical importance.

No collisions occur if the drop radius is 18 μ or less, and above that value for a_1, E is an increasing function of a_1 for fixed a. The effect of altering the drop size on the equations of motion (3) is solely to alter the parameter I which is a measure of the inertias of the drop and droplet. If the inertias are increased, the accelerations of the drop and droplet are decreased so that their horizontal separation alters more slowly and they are more likely to collide so that E is increased. The inertias when the drop has a radius less than 18 μ are small enough to make collisions impossible.

Collisions do not occur for sufficiently small droplets for each drop size (or, if the conjecture of the previous paragraph is not accepted, E becomes negligibly small). This can be partially explained by the low inertia of the droplet relative to that of the drop which allows it to be swept round the drop. But this would suggest that the drop roughly followed the streamlines of the flow past the drop which would give a non-zero value of E because of the size of the droplet. The trajectories shown in Fig. 2 indicate that as the droplet approaches the drop its direction of motion is changed rapidly, particularly if the initial horizontal separation is small, and it moves round the drop keeping close to it. This effect shows the importance of finding the flow patterns accurately when the spheres are close together. At the larger values for a, it is also found that there is a limiting droplet size above which no collisions can occur. When the drop and the droplet are of nearly equal size, the forces causing the horizontal separation to increase are small but the difference in the vertical velocities is also small, so that the forces have a long time in which to produce the horizontal displacement which is sufficient to ensure that collision does not occur. The general features of Fig. 1 have now been explained.

The present results must now be compared with those previously obtained. The table of results given by Mason (1957) are those found when the drop is much larger than the droplet but some results are given with a as large as 0·3. The values of E given in that table for drop sizes which fall in the range considered here are all larger than those shown in Fig. 1. This is to be expected as they were calculated on the assumption that

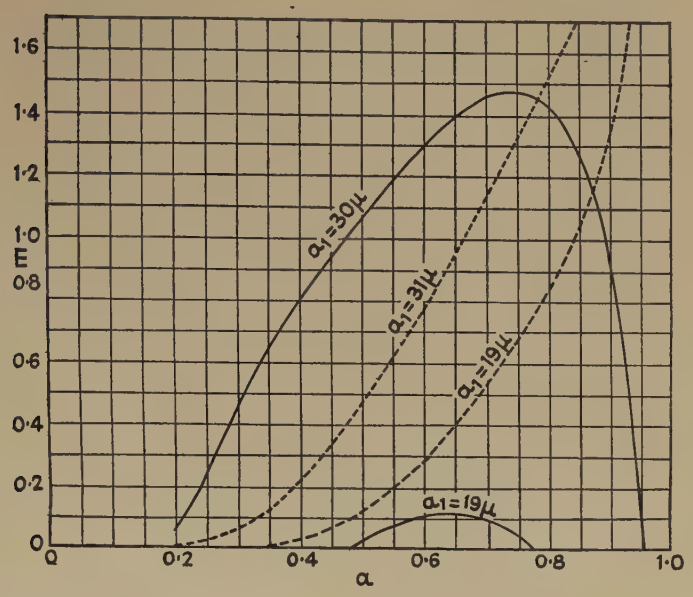

Figure 3. The collision efficiency of small drops. The results of the present calculations (full lines) are compared with those of Pearcey and Hill (broken lines).

the motion of the drop was unaffected by the presence of the droplet. When the acceleration of the drop is included, collision is less likely to occur.

The results found by Pearcey and Hill (1957) for the range up to 30 μ radius differ greatly from those found here. In the first place, they find collisions occurring for all drop sizes provided the droplet is large enough, whereas the present calculations give no collisions for drops of radius 18 μ or less. In agreement with the present results, they find that there is a limiting droplet size for each drop size below which collision cannot occur, but they do not find the limiting upper value for the droplet size. On the contrary, they find that E increases as the ratio of the radii tends to 1, taking values as high as 100 or more, and these large values of E they attribute to the influence of the wake. Values of E taken from Pearcey and Hill's paper for drops of radius 19 μ and 31 μ are shown in Fig. 3 with the present results for drops of radius 19 μ and 30 μ. The curve for drop radius 31 μ would lie slightly above that for radius 30 μ. The dissimilarity between the two pairs of curves emphasizes the importance of determining the forces on the spheres as accurately as possible, making full allowance for the mutual interference of the flows past the spheres.

The main differences between the Oseen and Stokes solutions at *low* Reynolds numbers occur close to the sphere, where the Stokes solution is more accurate, and at large distances from the sphere outside the wake; within the wake the two solutions are in close agreement. The wake can be defined as the vaguely paraboloidal region, including some fluid ahead of the sphere, outside which the vorticity is negligible. Defined in this way, the wake in the Stokes solution occupies the whole field of flow. In the region where the Oseen solution is to be preferred, the velocities are small so that the effect on the collision efficiency is likely to be small. Thus the values of E calculated using the Oseen solution should be in close agreement at low Reynolds numbers with those calculated using the Stokes solution. The great difference between Pearcey and Hill's results and the present ones is probably to be explained by the more accurate estimate of the forces made in the latter. It is, however, possible that the asymmetry of the fluid motion may be effective in increasing E for values of a close to 1. Some model experiments by Schotland (1957) have shown that the asymmetry can produce collisions between drops of the same size but only if the initial vertical separation is sufficiently small. Although, as explained

later, quantitative results obtained from model experiments of the kind used by Schotland are suspect, it is probable that the limiting vertical separation for collisions to occur of 10 drop-diameters will be roughly the same for water drops in air, so that no collisions will occur between drops of equal size with the initial vertical separation of 25 diameters used in the present calculations.

In an attempt to find experimental values for E, some model experiments were conducted by Sartor (1954) and more accurately by Schotland and Kaplin (1956). By dropping steel spheres in a viscous sugar solution, it was possible to model all the physical quantities except the ratio of the density of the spheres to that of the surrounding medium, which is very large for water drops in air but was about six in the model experiments. The results found by Schotland and Kaplin gave values for E for various drop sizes up to 10·5 μ radius. The results found in this paper make E zero over the whole range of these experiments. The reason for the discrepancy is the difference in the ratio of the densities in the actual problem and the model experiments and is connected with the neglect of the unsteadiness of the fluid motion.

In all calculations of the trajectories of particles which are small enough for their motion to be governed by the Stokes equations, it is assumed that the time derivative of the fluid velocity can be neglected. But as the particle accelerates, so does the fluid, causing the fluid motion to be unsteady. Suppose that the particle has a characteristic length l, velocity U and density σ and, for simplicity, that it is accelerating in fluid at rest. If p and $\mathbf{q}$ are the pressure and fluid velocity and if the fluid has a density ρ and viscosity η, the Stokes equations are

$$\rho \frac{\partial \mathbf{q}}{\partial t} = -\nabla p + \eta \nabla^2 \mathbf{q} \quad . \quad . \quad . \quad . \quad (5)$$

The force on the particle is of order ηUl and as this equals the mass-acceleration of the particle, its acceleration is of order $\eta U/\sigma l^2$. But this must be of the same order as the acceleration term $\partial \mathbf{q}/\partial t$ so that the left-hand side of Eq. (5) is of order $\rho\eta U/\sigma l^2$; the right-hand side is of order $\eta U/l^2$. The time derivative term in the equations of fluid motion can therefore be neglected if ρ/σ is small, which is certainly the case for water drops in air but was not so in the model experiments of Schotland and Kaplin. As the equations governing the motion are different in the two cases, it follows that the experimental results give no reliable information about E for water drops in air. As it would seem to be impossible to model the ratio of the densities and also to produce small velocities with large spheres, model experiments will yield no quantitative information about the collision efficiencies of small drops. The importance of the parameter ρ/σ has also been shown by Pearcey and Hill (1956) who solve the equation of motion of an accelerating sphere, using Oseen's equations for the fluid motion (I am indebted to a referee for this reference).

Acknowledgments

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Soil temperatures near Nairobi 1954-1955

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Summary

Continuous records of soil temperature at Muguga, Kenya, (1° 13′S, 36° 38′E) under bare soil at nominal depths of one, three and six inches and one, two and three feet were taken for the period 1954-1955.

The form of the temperature wave proved remarkably constant from day to day and the diurnal variation was measurable down to about one foot. The mean of maximum and minimum temperature is a good approximation to the daily mean temperature, as obtained by numerical integration, at depths of 6 in. and over. Nearer the surface, however, the mean of maximum and minimum gives an overestimate of the daily mean temperature. The annual soil temperature wave was not sufficiently marked to justify a mathematical analysis, but the mean diurnal wave for selected ten-day periods was analysed and values of thermal diffusivity were computed for the depth ranges 1-6 in. and 6-14 in. Despite seasonal changes in soil moisture content, diffusivity values proved remarkably constant with mean values of 0·0020 and 0·0015 $cm^2\ sec^{-1}$ respectively for the two depth ranges. There was no obvious correlation with soil water changes.

1. Introduction

Six Negretti and Zambra mercury-in-steel distant recording soil thermographs were buried at depths of 1, 3 and 6 in. and 1, 2 and 3 ft under bare soil in the meteorological enclosure at Muguga, Kenya. Since the recording bulbs are 19 mm in diameter, accurate placing at a point was impossible, and it has been assumed that the temperature recorded is equal to that of the soil at the level of the mid-point of the bulb. There is, of course, a phase lag in the temperature observation, both in heating and in cooling, owing to the high conductivity and heat capacity of the bulb as compared with the soil which it displaces. Indeed, the instrument acts as a heat source and sink capable of causing considerable modification to the flow of heat in its immediate vicinity. For this reason, the thermometers were spaced over a horizontal distance of about six feet, instead of down a vertical profile, for the error involved in assuming that isotherms are planes parallel to the soil surface is likely to be less than that owing to the linear and thermal dimensions of the instruments, if placed in close proximity. The installation was completed in 1953, and the depths were checked and found correct three months later. A further check in 1957 indicated that the instruments at 3 in. and at 1 ft depth had risen 1 in. and fallen 1·75 in. respectively. Hence for calculation of the thermal constants of the soil, data from the 3 in. nominal depth have had to be neglected and the 6-12 in. depth range has been amended to the actual measured range of 6-13·75 in.

2. Diurnal and annual variations

In Fig. 1 a typical series of temperature records at all six depths is shown for a period of 5 days. The diurnal variation of soil temperature is not measurable with the present equipment at depths beyond 1 ft, and it is clear that even at about one-foot depth the daily amplitude is of the same order as the limit of accuracy of the measurements. Owing to the constant day length experienced at equatorial latitudes and the uniform variation of soil temperature from one day to the next, the form of the waves is remarkably constant. Contrary to the findings, under temperate conditions, of Keen and Russell (1921) and of Coutts (1955), calculation of daily mean temperatures over ten 10-day periods in 1954 and 1955, by averaging the daily maximum and minimum, did not result in good estimates of the true daily mean temperature. In fact the temperature at 1 in. depth was overestimated by 5°F, and at about 3 in. depth by 1·5°F. For depths of 6 in. and beyond, the error is insignificant.

In Fig. 2 are plotted monthly mean values for 1955 of soil temperature daily maxima at 1 in., 3 in. and 6 in. and of daily mean soil temperatures at 1 ft and 2 ft, together with the maximum screen air temperature and the total solar radiation incident on a horizontal surface at the upper limit of the atmosphere at latitudes 0° and 50°N. It is clear that equatorial soil- and air-temperature

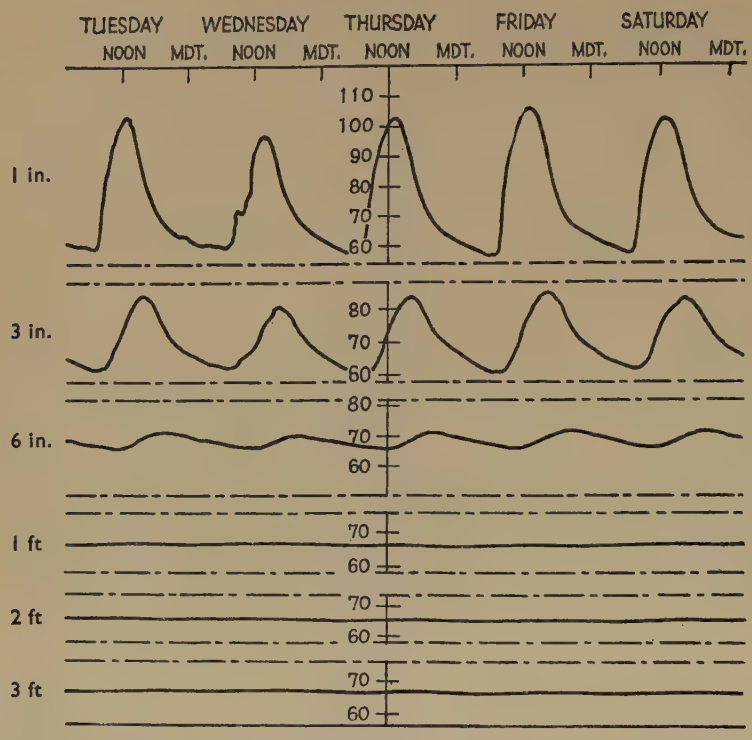

Figure 1. Muguga soil temperature records, 6 to 10 Dec. 1955.

variations are not dominated by seasonal variations in solar latitude, as in temperate latitudes, and it is therefore pointless to study the penetration of the annual-temperature wave, on the assumption of a sinusoidal forcing wave applied at the surface.

Attention was therefore directed to the progression of the diurnal wave into the soil, using records from the instruments placed at depths down to 13·75 in. From consideration of the available meteorological and soil moisture data, a series of nine 10-day periods was chosen as representative of all stages in the moisture cycle of the soil. Two-hourly mean values for each of these ten-day periods were calculated and a Fourier analysis for the first and second harmonics was found to approximate very closely to the observed data. In Fig. 3 curves of this diurnal variation at nominal depths of 1, 3 and 6 in. are presented for the periods 11-20 January 1955 and 11-20 July 1955, and the residuals at each depth are plotted on the same scale.

The Fourier analysis provided values of the amplitude θ, and phase ψ of the soil temperature wave at each depth x. From these the thermal parameters a and b can be calculated using the relations

$$a = \frac{\log_e \theta_1/\theta_2}{x_2 - x_1} = \frac{\alpha}{x_2 - x_1} \quad . \quad . \quad . \quad . \quad (1)$$

$$b = \frac{\psi_2 - \psi_1}{x_2 - x_1} = \frac{\beta}{x_2 - x_1} \quad . \quad . \quad . \quad . \quad (2)$$

where α and β are defined by the above equations as the logarithmic decrement and the phase lag respectively of the soil temperature wave over the depth range $x_2 - x_1$.

The thermal diffusivity of the soil K over the range $(x_2 - x_1)$ is then given by the relation (McCulloch and Penman 1956) :

$$K = \frac{7{\cdot}27 \times 10^{-5}\, a}{b\,(a^2 + b^2)} \text{ for the fundamental} \quad . \quad . \quad . \quad (3)$$

Eq. (3) is an approximation to a more comprehensive equation, the use of which was not possible owing to the small number of thermometers down the profile.

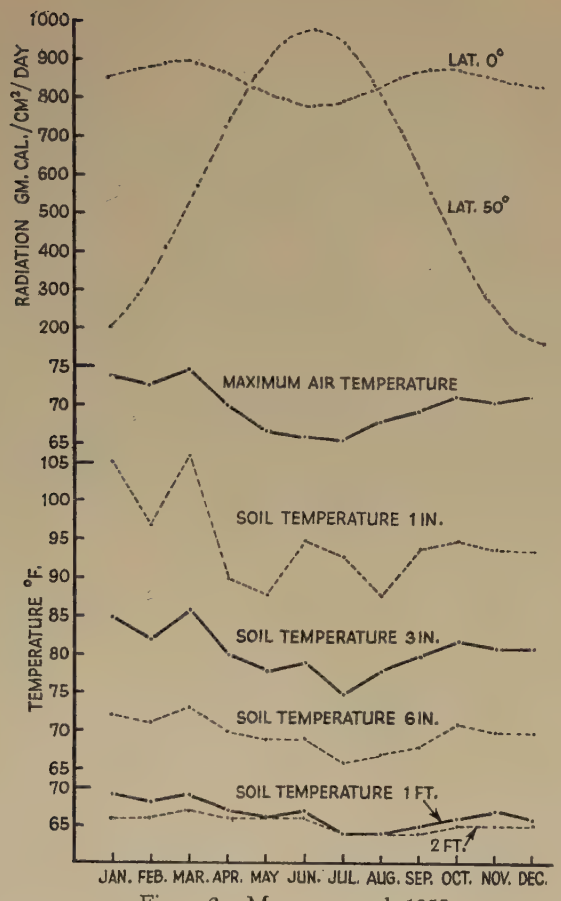

Figure 2. Muguga records 1955.

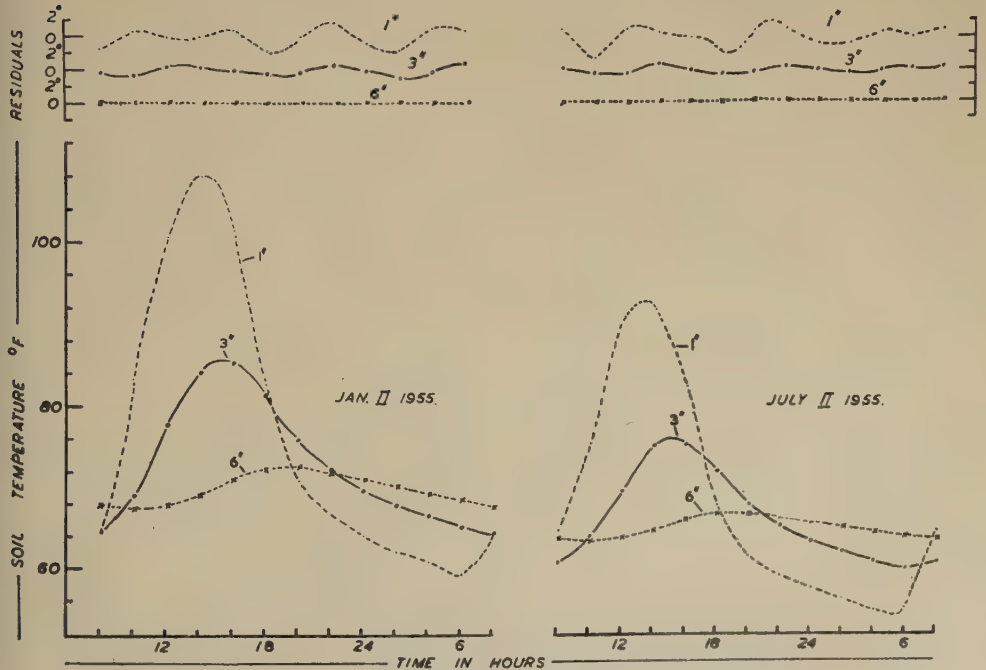

Figure 3. Diurnal variation of soil temperature.

3. Effect of the soil moisture regime on the thermal parameters

Because of the movement of two of the thermometers, and the consequent doubts as to the depth ranges for which values of the thermal constants of the soil can be calculated, a study of moisture effects was made over the nine 10-day periods, using values of α and β for the fundamental and second harmonic, calculated for the nominal depth ranges 1-3 in., 3-6 in. and 6-13·75 in. (Table 1). The accuracy of these actual ranges is quite irrelevant for the present purpose, provided that no variation in depth occurred during the period of the analysis November 1954-July 1955. This is a reasonable assumption to make, since the maximum amount of soil movement would clearly have occurred during the first 12 months after installation of the instruments. Standard errors were calculated from the variation of the original data on the assumption that the reliability of each two-hourly mean value was the same. Weighting of the values of α and β according to standard error for each of the two harmonics resulted in a weighted mean value which differed little from that derived by use of the fundamental alone, and so data for the fundamental only were used in calculations of thermal diffusivity. It is felt, however, that the good agreement, seldom seen in other published records of soil temperature, between values derived from both harmonics justifies the presentation of the inevitably less precise data for the second harmonic. In Table 1 the column entitled soil moisture gives a general indication from soil moisture sampling on an experimental plot nearby, at depths 1-6 in. and at 1 ft, of the moisture status of the soil during the period.

TABLE 1. Values of α and β, for fundamental and second harmonic

Period		α		β		Soil moisture
		Fundamental	Second harmonic	Fundamental	Second harmonic	
		(a) Depth range 1-3 in. nominal				
November	I	0·52 ± 0·06	0·46 ± 0·16	0·41 ± 0·06	0·37 ± 0·16	Wilting point
	II	0·56 ± 0·09	0·50 ± 0·27	0·47 ± 0·09	0·46 ± 0·27	Moist
January	I	0·78 ± 0·05	0·68 ± 0·14	0·55 ± 0·05	0·54 ± 0·14	Wilting point
	II	0·86 ± 0·03	0·71 ± 0·07	0·53 ± 0·03	0·44 ± 0·07	Wilting point
February	I	0·56 ± 0·06	0·49 ± 0·16	0·58 ± 0·06	0·60 ± 0·16	Moist
	II	0·64 ± 0·04	0·54 ± 0·11	0·56 ± 0·04	0·54 ± 0·11	Wet
April	I	0·62 ± 0·04	0·53 ± 0·11	0·57 ± 0·04	0·55 ± 0·11	Wilting point
	II	0·51 ± 0·07	0·47 ± 0·11	0·57 ± 0·07	0·64 ± 0·11	Wet
July	II	0·88 ± 0·05	0·79 ± 0·15	0·65 ± 0·05	0·63 ± 0·15	Wilting point
		(b) Depth range 3-6 in. nominal				
November	I	1·46 ± 0·09	1·46 ± 0·46	1·15 ± 0·09	1·36 ± 0·46	Wilting point
	II	1·34 ± 0·20	1·48 ± 1·23	1·18 ± 0·20	1·39 ± 1·23	Moist
January	I	1·43 ± 0·06	1·40 ± 0·32	1·17 ± 0·06	1·27 ± 0·32	Wilting point
	II	1·35 ± 0·04	1·30 ± 0·17	1·14 ± 0·04	1·06 ± 0·17	Wilting point
February	I	1·22 ± 0·09	1·22 ± 0·37	1·16 ± 0·09	1·10 ± 0·37	Moist
	II	1·30 ± 0·08	1·21 ± 0·34	1·18 ± 0·08	1·16 ± 0·34	Wet
April	I	1·21 ± 0·06	1·14 ± 0·24	1·14 ± 0·06	1·10 ± 0·24	Wilting point
	II	1·28 ± 0·15	1·35 ± 0·76	1·12 ± 0·15	1·27 ± 0·76	Wet
July	II	1·42 ± 0·08	1·32 ± 0·36	1·09 ± 0·08	1·22 ± 0·36	Wilting point
		(c) Depth range 6-13·75 in.				
November	I	2·09 ± 0·36	1·63 ± 2·46	3·42 ± 0·36	1·64 ± 2·46	Wilting point
	II	1·55 ± 0·92	1·06 ± 5·68	3·66 ± 0·92	2·30 ± 5·68	Wilting point
January	I	1·58 ± 0·20	1·17 ± 1·16	3·21 ± 0·20	2·58 ± 1·16	Wilting point
	II	2·36 ± 0·36	1·17 ± 0·85	2·94 ± 0·36	3·06 ± 0·85	Wilting point
February	I	2·39 ± 0·87	1·05 ± 1·62	3·35 ± 0·87	2·39 ± 1·62	Wilting point
	II	2·57 ± 0·55	1·12 ± 0·95	2·54 ± 0·55	3·02 ± 0·95	Moist
April	I	2·23 ± 0·28	1·46 ± 1·06	2·59 ± 0·28	2·65 ± 1·06	Wilting point
	II	2·40 ± 0·52	1·18 ± 4·11	1·95 ± 0·52	2·22 ± 4·11	Wet
July	II	2·34 ± 0·50	1·42 ± 1·82	3·26 ± 0·50	2·39 ± 1·82	Wilting point

In Table 1 (a) and (b) the actual values of α and β from 1-6 in. derived from both harmonics, lie substantially within the limits set by the quoted standard errors. Some variation occurs between periods but no correlation with soil moisture changes can be detected. In general also, α is numerically greater than β and the range of variation of α from one period to another exceeds that of β. Table 1 (c) presents data for the 6-13·75 in. range. Results for this range are less reliable, because of observational errors due to the very small amplitude of the diurnal variation at 1 ft depth. Again no clear dependence on soil moisture content can be discovered.

4. Calculation of thermal diffusivity

Data for the fundamental temperature wave at 1 in., 6 in. and 13·75 in. were used to calculate values of a, b and hence K, according to Eq. (3), and the results are tabulated in Table 2 (a) and (b) respectively. In the range 1-6 in. values of a varied somewhat, though values of b and hence K were effectively constant with a mean value of thermal diffusivity of 0·0020 $cm^2\ sec^{-1}$. No dependence on moisture content is discernible. For the 6-13·75 in. layer, values of a and b varied more widely, and resultant values of diffusivity ranged from 0·0010 to 0·0022 $cm^2\ sec^{-1}$ without any dependence on soil moisture changes. While this variation is well within the bounds of physical possibility, it is equally accountable on the basis of errors of estimation of α and β.

TABLE 2. Values of parameters a and b derived from the fundamental, with values of K computed from them

Period			a	b	K c.g.s. units	Soil moisture
			(a) Depth range (actual) 1-6 in.			
November	I	1954	0·156	0·123	23×10^{-4}	Wilting point
	II	1954	0·150	0·133	20×10^{-4}	Moist
January	I	1955	0·174	0·136	19×10^{-4}	Wilting point
	II	1955	0·174	0·131	20×10^{-4}	Wilting point
February	I	1955	0·140	0·137	19×10^{-4}	Moist
	II	1955	0·152	0·137	19×10^{-4}	Wet
April	I	1955	0·144	0·134	20×10^{-4}	Wilting point
	II	1955	0·142	0·133	21×10^{-4}	Wet
July	II	1955	0·172	0·137	19×10^{-4}	Wilting point
			(b) Depth range (actual) 6-13·75 in.			
November	I	1954	0·106	0·174	11×10^{-4}	Wilting point
	II	1954	0·240	0·186	10×10^{-4}	Wilting point
January	I	1955	0·080	0·163	11×10^{-4}	Wilting point
	II	1955	0·120	0·149	16×10^{-4}	Wilting point
February	I	1955	0·122	0·170	12×10^{-4}	Wilting point
	I	1955	0·130	0·129	22×10^{-4}	Moist
April	I	1955	0·113	0·132	21×10^{-4}	Wilting point
	II	1955	0·122	0·137	19×10^{-4}	Wet
July	II	1955	0·119	0·165	13×10^{-4}	Wilting point

These results suggest that while the mercury-in-steel thermometers were not suitable for measurement of the small changes of temperature at 1 ft and beyond, they did serve a useful purpose nearer the surface in smoothing out the inevitable small random and unaccountable variations of the temperature of the soil. The consistency of the values of K in Table 2 (a) certainly encourages confidence in the method of handling the data.

5. Discussion

These results are comparable with the determinations of Smith (1939) using Barns loam at different moisture contents, from 5-27 per cent. A few selected values of the physical properties of both soils are quoted in Table 3. The Barns loam is clearly a much more sandy material than

the highly porous lateritic Kikuyu red loam at Muguga, and the moisture characteristics of these soils differ markedly. Thus even at wilting point, the moisture content of the Kikuyu clay loam is over 20 per cent and so this particular soil is never really dried out in the field, except possibly in the first inch depth. In both cases, however, the thermal diffusivity showed a negligible dependence on moisture content.

TABLE 3

Soil	Depth (inches)	Particle size (per cent) Sand	Slit	Clay	Bulk density g/cc	Moisture content experimental range (per cent)	Mean K cm^2 sec^{-1}
Kikuyu red loam	1-6	13	13	74	0·90	approx. 18-42	0·0020
	6-14					21-36	0·0015
Barns loam		42	25	33	1·16	5-27	0·0014

Patten (1909) in his classical work on the dependence of the thermal constants of the soil on moisture content, showed, using artificially packed soil columns, that the thermal diffusivity of a soil varies much less with moisture content than does the thermal conductivity. A physical explanation of this result has been given by Keen (1931). De Vries (1952) has developed a method for estimating soil moisture from measurements of thermal conductivity but has shown that the method is applicable only to soils for which the relationship between thermal conductivity and volume fraction of water is steep over the moisture range of importance in the field to field capacity to wilting point. (' This will generally be the case with sandy soils,' de Vries 1953). In view of the high clay content of the Muguga soil, it is felt that this may well account for the almost complete independence of the thermal diffusivity on moisture content which has been observed at Muguga.

Considerable transport of water vapour under temperature gradients in the soil was observed by Hursh and Pereira (1953) in Kenya. Such transport (de Vries 1950) has shown will introduce an additional term in the general equation for heat flow. Values of diffusivity observed, then, comprise two components – the true soil thermal diffusivity plus a term due to vapour transport. Using the physical data for the Muguga soil, at field capacity and at wilting point, the size of the vapour contribution to apparent thermal diffusivity has been computed at temperatures of 68°F (20°C) and at 104°F (40°C), by the method of de Vries, (Table 4). Clearly this term is only of the order of 5 to 10 per cent of the measured diffusivity, and so cannot be detected in this present work.

TABLE 4. Data for Kikuyu red loam

Moisture status	Air filled pore space Total volume per cent	Sp. heat/unit vol. cal/cc/°C	Vapour contribution to K cm^2 sec^{-1} Temperature 68°F	Temperature 104°F
Field capacity	30	0·70	$0{\cdot}5 \times 10^{-4}$	$1{\cdot}6 \times 10^{-4}$
Wilting point	45	0·50	$1{\cdot}1 \times 10^{-4}$	$3{\cdot}3 \times 10^{-4}$

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551.573

Observable evapotranspiration in the Basin of the River Thames

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Summary

The measurement of evapotranspiration from large land areas is, like the measurement of evaporation from large areas of water, a formidable problem and not often accurately accomplished. Nevertheless, it is essential for an engineer to try to check the results of theoretical or empirical analyses, or small-scale experiments, against nature where natural phenomena are concerned and the results of an analysis of the hydrology of the Thames are presented herein. Conclusions favouring the Thornthwaite method are reached.

1. Introduction

Of the theoretical assessments of potential evapotranspiration in Britain, those of Penman are most frequently quoted and are based on his energy-balance thesis. For example, Penman (1951) gives the potential evapotranspiration of the Thames Basin above Teddington per annum, as ranging between 17 in. and 21 in. In support he quotes the accepted fact that the long-term difference between rainfall and run-off in the Thames Basin is about 19 in. per annum. In addition, Penman exhibits estimated changes in storage in three chalk wells near Winchester, Odiham and Purley, for a term of 20 years from 1930 to 1949. He retains (Penman 1955) his earlier assessment of potential evapotranspiration and also indicates his assumption that in the Hursley Well, near Winchester, a 20 ft rise in well-level corresponds to 2 in. of rain stored.

An alternative approach is that of Thornthwaite, of which a paper by Howe (1956) is a recent reconsideration. In this, the potential evapotranspiration of the Thames Basin is given as about 25 in. per annum.

Several workers have carried out relatively small-scale experiments into water losses from land surfaces and plants with a variety of results. Rider (1957) concludes that Penman's fundamental general assumption, that the potential transpiration is the same as that from grass, provided that the crop has the same colour as grass and completely covers the ground, should be approached with the utmost caution. In addition, Rider quotes actual observed values of various crops. Stanhill (1958) adduces data on observed evapotranspiration from short grass at Warwick, of the order of 26 in. per annum.

It would be difficult to find another basin about which more is known. Authoritative rainfall and run-off data have been published covering at least 71 years (Stock 1950; H.M.S.O. *Surface-water Year Book* (various)). The geology, geography and climate of its catchment are so well known as to require little further definition. Glasspoole (1951) has given a useful summary of Thames data and references. Gold (1951) has propounded a modified empirical method for predicting the Thames run-off from the rainfall and reports a number of anomalies both in this and in other similar computations.

2. Hydrological balance

There are, nevertheless, two major difficulties in deducing the actual and potential evapotranspirations from this wealth of data and reference. They are (a) what proportion of the rainfall is temporarily stored underground in wet years and (b) what proportion of the rainfall escapes underground out of the basin without being measured ?

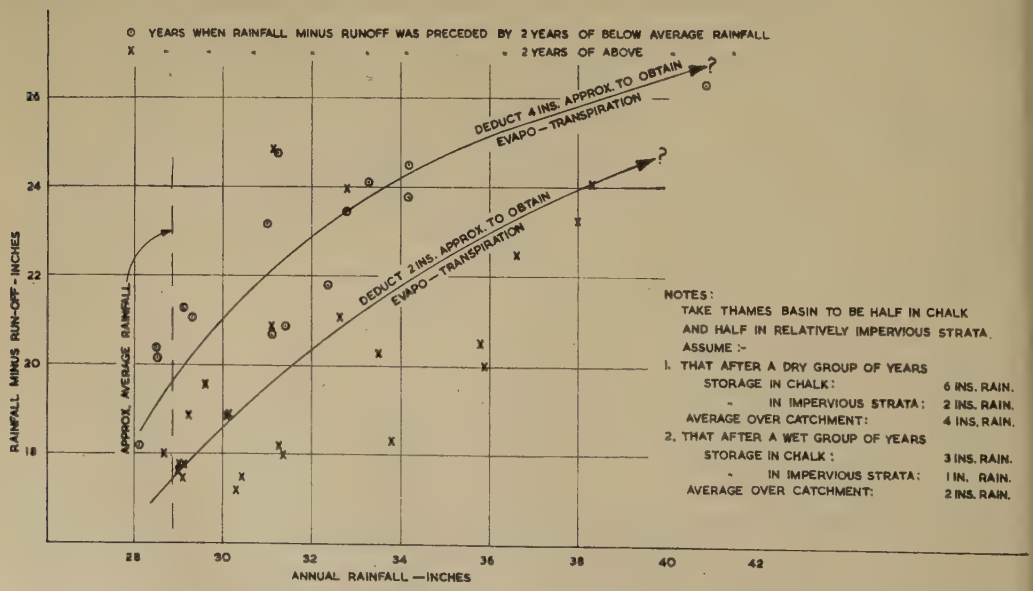
From the rapidity with which rain disappears into chalk which occupies a substantial proportion of the Thames catchment, it might be thought that a very large proportion of the rainfall would have to be considered as lost beyond measurement, but this would be erroneous. Much of the rain which sinks into the chalk hills is held up by capillary action or reappears again quite rapidly in springs at their bases, to return to the rivers. Only a limited proportion of the water in the chalk is free, much being adsorbed and more or less fixed. A recent paper (Croney, Coleman and Bridge 1952) indicates the enormous forces which are required to dislodge such water.

Not unexpectedly, this subject has been much discussed by engineers and many papers have been published over more than a century. Lapworth (1948) gives much invaluable data. It is essential to understand that by percolation, he means water which sinks into upland-chalk surface and not water which is entirely stored underground. On the basis of the levels of five wells he concludes that the storage equivalent to the extreme range of fluctuation over many years of record, varies from 12 in. to 18·1 in. with an average of 14 in. He notes that the highest and lowest levels may occur at times which are many years apart.

Of more specific intention, the work of Halton Thomson at the Compton Well is now almost a classic. Reference may be made to his paper (1947) and this work is of particular value because the well is unpumped and the water-levels represent a true water-balance. It is of interest to note that the greatest increase in storage in any one year in the chalk was equivalent to about 5½ in. of rainfall, which is very much in keeping with the results from the wells referred to by Lapworth and Penman.

If we assume that, after a period of less than average rainfall, chalk and similar lands could retain unused, in store, in a year, an additional 6 in. of rainfall, we would probably be making a very ample allowance. The other areas of the catchment are much less pervious and might be able to retain another 2 in. of rain, unused, in store. Since the proportionate areas are about half-and-half, we may reasonably allocate 4 in. overall in such circumstances. After a period of greater than average rainfall, an overall average of 2 in. retention unused would be reasonable.

As regards the water passing out of the Basin unmeasured at Teddington Weir, the London ' artesian ' Basin is a fair indicator because it has a large determinate pervious catchment including much chalk. Over a century ago there were great hopes that London's water supply might be drawn from wells from this great basin, but the controversy became somewhat academic with the publication of the Rev. J. C. Clutterbuck's paper in 1850. The essence of this paper and its discussion was that an abstraction of some 12,000,000 gallons per day (22·3 cusecs) at that time, was reducing the water-table very appreciably. Since then the water-table has been drawn down to a great extent (Wilson and Grace 1942). Considering the area of pervious land feeding this basin we can only say that a subterranean aquifer incapable of sustaining a mere 22 cusecs is not very significant when compared with a Thames whose average flow is about 2,675 cusecs. Since there are no geological peculiarities in the Thames Basin which would vitiate reference to the behaviour of the London Basin, we may conclude that external leakage is relatively small. London also draws some water from wells in strata above the chalk (H.M.S.O. 1938), but we are only concerned with such water as may seep past Teddington Weir unmeasured and this cannot be very significant because of the strata and the small hydraulic gradient.


⊙ YEARS WHEN RAINFALL MINUS RUNOFF WAS PRECEDED BY 2 YEARS OF BELOW AVERAGE RAINFALL
× " " " " " " 2 YEARS OF ABOVE " "
DEDUCT 4 INS. APPROX. TO OBTAIN EVAPO – TRANSPIRATION
DEDUCT 2 INS. APPROX. TO OBTAIN EVAPO – TRANSPIRATION
APPROX. AVERAGE RAINFALL
RAINFALL MINUS RUN-OFF - INCHES
ANNUAL RAINFALL — INCHES
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NOTES:
TAKE THAMES BASIN TO BE HALF IN CHALK AND HALF IN RELATIVELY IMPERVIOUS STRATA.
ASSUME :-
1. THAT AFTER A DRY GROUP OF YEARS
STORAGE IN CHALK: 6 INS. RAIN.
" IN IMPERVIOUS STRATA: 2 INS. RAIN.
AVERAGE OVER CATCHMENT: 4 INS. RAIN.
2. THAT AFTER A WET GROUP OF YEARS
STORAGE IN CHALK: 3 INS. RAIN.
" IN IMPERVIOUS STRATA: 1 IN. RAIN.
AVERAGE OVER CATCHMENT: 2 INS. RAIN.

Figure 1. Evapotranspiration in the Thames Basin.

The mean annual rainfall on the Thames catchment of 3,812 sq. mi, over a period of 71 years, s 28·9 in. The mean annual natural river flow to Teddington is equivalent to 9·6 in. Since neither the long-term storage nor subterranean escape can be relatively very significant we take t that the actual long-average evapotranspiration is about 19·3 in. per annum.

3. Potential evapotranspiration

As regards the potential evapotranspiration, the position may be somewhat different, for in 32 out of the 71 years of record, the difference between rainfall and run-off was greater than 19 in.; once it was over 26 in. per annum.

Since this analysis is concerned with potential evapotranspiration, i.e., evaporation and transpiration when there are no deficiencies in the amount of water present to be evaporated and transpired, we may consider these 32 years in more detail. On Fig. 1 are plotted individually, the rainfalls and ' differences ' and a distinction is made between years preceded by a two-year period of less-than-average rainfall and those preceded by a two-year period of greater-than-average rainfall. In the former cases we may expect more water to go into temporary storage than in the latter cases. Probably in the latter cases water would be drawn from, rather than added to storage, but that will not be assumed at present. Taking 4 in. overall to storage in the former and 2 in. in the latter case, trend lines are drawn and several conclusions seem apparent;

(1) Evapotranspiration in years of heavy rainfall, following a period of less-than-average rain, has observably exceeded 22 in. per annum on a still rising curve.

(2) Evapotranspiration in years of heavy rainfall, following a period of greater-than-average rain, has observably exceeded 22 in. per annum on a still more rapidly rising curve.

(3) The difference between rainfall and run-off modified to allow for storage, and thus a measure of evapotranspiration, rises with the rainfall and has not reached a distinguishable maximum in the 71 years on record.

It is relevant to point out that many of these years were of greater-than-average cloudiness, so that incident solar energy can hardly have been above average.

From the trend of the observations it is hardly possible to escape the conclusions on hydrological grounds that average potential annual evapotranspiration in the Thames Basin is at least in the middle or upper twenties of inches. This supports Thornthwaite-type results rather than Penman's and is supported by experimental data such as Stanhill's.

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551.594.11 : 551.594.25

Measurements of the atmospheric potential gradient on a Canadian glacier

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(Manuscript received 23 June 1958)

SUMMARY

Measurements of the atmospheric potential gradient made on a glacier at an altitude of 1,700 m showed a behaviour considerably simpler than is normally found near the ground. The nearness of the cloud, and the absence of point discharge for some kilometres gave a clear-cut distinction between snow and rain, these almost always being associated with positive and negative fields respectively. The results suggest that the electrification processes are associated with the precipitation in both cases, and that marked electrical effects accompany the melting of snow.

1. INTRODUCTION

During the summer of 1957 the author took part in an expedition to the Salmon Glacier in British Columbia, and it was decided to take equipment for making measurements of the atmospheric potential gradient.

The glacier, which is situated in the Canadian coastal range of mountains just outside the southern tip of Alaska (56° 10′N, 130° 7′W), is some twelve kilometres long, descending from snowfields at an altitude of 1,700 m to the snout at 210 m above sea level. The work there included a study of the meteorology of the accumulation area (Adkins 1958) and for this purpose a camp was established early in June at the top of the glacier. From the camp, the snowfields rose another 300 m to the north, while in other directions they descended into other glaciers. Surrounding mountain peaks, the nearest about four kilometres away, rose to 2,400 m.

The camp provided an unusual situation for electrical measurements. Being in the centre of a large snow-covered plateau, the effects of point discharge were negligible, and the altitude was such that the camp was often in cloud or near the cloud base. Thus it might be expected that some of the complications which are normally associated with measurements from the ground would be removed.

2. APPARATUS

It was necessary for the equipment to be simple, portable and of low power requirements. It was therefore decided to use a polonium radioactive collector and a small electrometer unit.

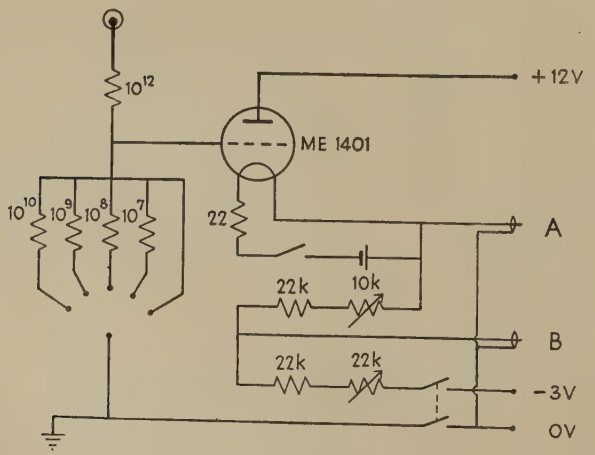

Figure 1. Electrometer circuit.

The electrometer circuit is shown in Fig. 1. A typical new polonium collector gives an effective source resistance of about $2 \times 10^{10}\ \Omega$ (Scrase 1934), so that the electrometer with its input impedance of $10^{12}\ \Omega$ effectively measures the collector potential. The valve operates essentially as a cathode follower providing a voltage output, *A*, and a current output, *B*. The sensitivity of the current output is controlled by the 10 kΩ resistor in the cathode and a bucking circuit taken to -3 v brings the current to zero for zero field. The ME 1401 has a stated maximum grid current of about 10^{-14} a, so that on the most sensitive range the unit is stable to better than 0·1 per cent in this respect. In practice, a small zero drift of a few per cent often took place during a series of observations owing to battery changes. For this reason the zero was checked at the end of each run and correction made.

The circuit, with batteries and desiccator, was built into a small duralumin case, switches and connectors being brought out to one side. The whole unit was sealed, but for access to the batteries and desiccator the base could be unscrewed, the seal here being made with a rubber washer. The source was supported on a brass rod which plugged into a connector in a large polythene insulator fixed to the top of the unit. A polythene shield protected the top and sides of the insulator from precipitation.

During use the unit was placed on the snow, well away from the camp and connection was made to a meter in the camp where readings were taken at minute or half-minute intervals. The results were normally plotted directly and notes on meteorological conditions added to one side of the graph. Some of the records are reproduced in the examples of § 3. Local time is used throughout.

Unfortunately, the source and part of its support had to be abandoned on the glacier and it was not possible to make a detailed calibration of the device. The fields quoted are obtained by taking the fine-weather field as 110 v m^{-1}, which seems a reasonable value for the place of observation.

3. Results

There was a sharp distinction between the electrical effects of snow and rain. In the former the potential gradient was virtually always positive, while in the latter it was virtually always negative, and in both cases larger rates of precipitation were associated with larger fields. On some occasions during rain the correlation was very close. No observations were made during thunderstorms, but the negative potential gradients of rain persisted in quite violent conditions. In drizzle, the field was usually small and of either sign.

On the occasion shown in Fig. 2, the camp was in cloud which was probably no more than a few hundred metres thick as the sun could often be seen through it. Intermittent drizzle fell for some hours and it can be seen that during nearly all the period of observation the field remained positive, though less than that of fine weather. At 1518 LMT the cloud became thicker and a few drops of rain fell. During the next five minutes large drops continued to fall and the field became negative, reaching about -100 v m^{-1}. It is possible that the ' drops ' were partly melted snow flakes. With the return to drizzle the field soon became positive again and continued so throughout the rest of the period.

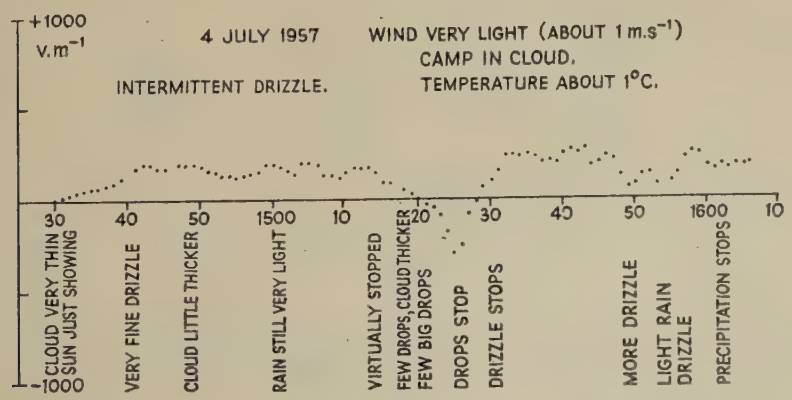

Figure 2. Potential gradient, 4 July 1957.

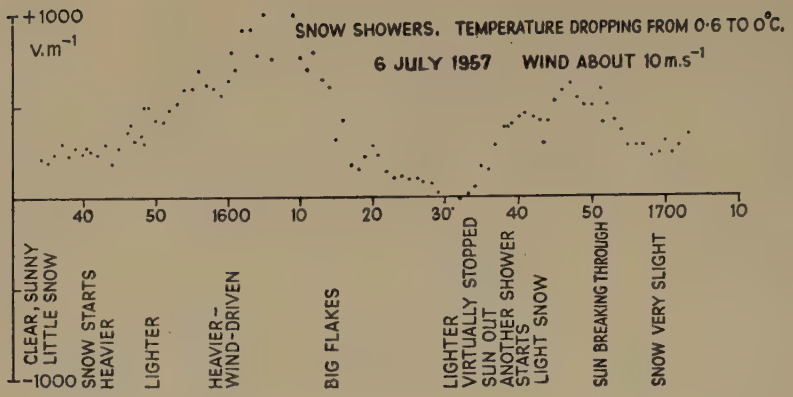

Figure 3. Potential gradient, 6 July 1957.

On 6 July isolated snow showers occurred, the snow falling from small cloud systems between which the weather was bright and sunny. A period during which two showers passed is shown in Fig. 3. At the beginning of the observations the sky above the camp was clear and the field near the fine-weather value. During the first shower the field increased, reaching over + 300 v m^{-1} during the period of heaviest snow. Towards the end the field decreased and finally became negative (about − 20 v m^{-1}) for about two minutes. The second shower followed the first closely and again the snow was accompanied by an increased positive field. If we suppose that the snow carried a negative charge, as has been found by other observers, and that the transport of this charge to earth accounts for the increased positive fields, then the period of negative field is simply explained in terms of wind shear. The negative snow trails behind the cloud, and for the short period during which the cloud has passed and snow is still falling, the field is determined by the charge on the snow, which here is sufficient to cause the field to become negative.

Fig. 4 shows two hours of observations during prolonged showery rain. It can be seen that the field is negative throughout, and that it sometimes becomes very big. It should be noticed how the largest fields are associated with the greatest rainfall. In particular it should be observed how closely the short periods of increased intensity between 1250 and 1330 LMT are correlated with increased negative fields. After the period of the example, drizzle continued for some time, the field remaining quiet at a little over − 100 v m^{-1}.

On 11 June it had been raining all the morning when at 1400 LMT the rain turned to snow. At 1515 LMT it was still snowing and field observations were begun. (Fig. 5). Until 1630 LMT the field remained positive, at times reaching + 100 v m^{-1} but averaging about + 30 v m^{-1}.

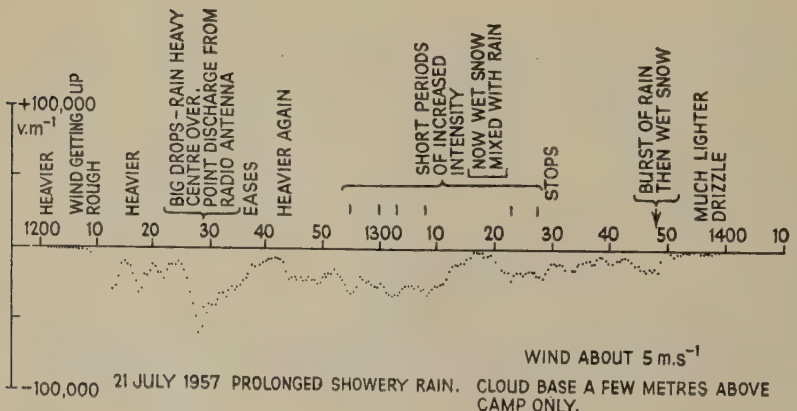
+100,000
v.m⁻¹
HEAVIER
WIND GETTING UP ROUGH
HEAVIER
BIG DROPS - RAIN HEAVY CENTRE OVER, POINT DISCHARGE FROM RADIO ANTENNA EASES
HEAVIER AGAIN
SHORT PERIODS OF INCREASED INTENSITY
NOW WET SNOW MIXED WITH RAIN
STOPS
BURST OF RAIN THEN WET SNOW
MUCH LIGHTER DRIZZLE
1200 10 20 30 40 50 1300 10 20 30 40 50 1400 10
WIND ABOUT 5 m.s⁻¹
-100,000
21 JULY 1957 PROLONGED SHOWERY RAIN. CLOUD BASE A FEW METRES ABOVE CAMP ONLY.

Figure 4. Potential gradient, 21 July 1957.

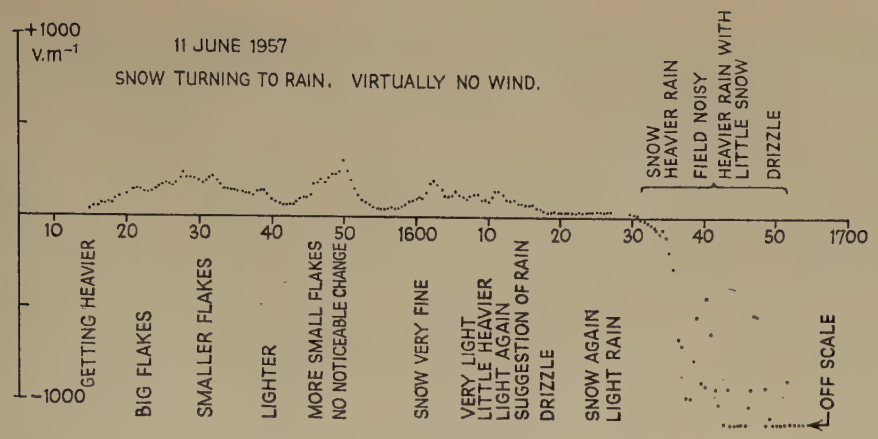

Figure 5. Potential gradient, 11 June 1957.

At about 1615 LMT the freezing level began to rise above the camp and the snow gradually gave way to drizzle. Soon after 1630 LMT it began to rain, the rain soon becoming moderately heavy, and alternate rain and drizzle continued for some time. The behaviour of the field at the transition to rain is of particular interest. Having remained positive throughout the snow, it quickly became negative and took on the ' noisy ' appearance which is associated with the presence of charge near the ground. After 1700 LMT the observations were continued for some 30 min with reduced sensitivity and showed fields reaching over $-$ 15,000 v m^{-1}. These results strongly suggest that accompanying the melting of snow there is a release of negative charge to the air giving positive rain and a negative field.

4. Discussion

The above results have a simple explanation if the cloud-electrification processes are associated with the precipitation charges. Snow carrying a negative charge to earth will give rise to positive fields, and rain carrying a positive charge to earth will result in negative fields. In both cases, increased precipitation rates will be associated with greater fields. The fact that the correspondence between precipitation rate and field is so readily observed in the present measurements is certainly the result of the site of observation. Apart from the absence of point discharge, the camp was always either in the cloud or near to the cloud base so that increased rainfall took place very close to the centres in which the rain was produced. We may deduce that rain leaves negative charge in these centres, and, further, we may be sure that the charge is shed low in the cloud, for otherwise the close correspondence would be lost.

In the case of snowfall it is difficult to be certain that the blowing of the snow by wind is not important. Certainly the occasions when the potential gradient reached particularly high values in snow were ones of considerable wind (during one period of severe weather the potential gradient reached over + 3,000 v m^{-1}). However, it does seem that separation of charge does take place in snow, although the associated electrical effects are bound to be smaller than in the case of rain as a result of the smaller terminal velocity of snowflakes.

We thus arrive at a picture where (a), positive charge is left in the air in regions producing snow, the snow removing the negative charge, and (b), melting snow releases negative charge to the air giving the rain a positive charge. The requirements that snow carry a negative charge and rain a positive are in general agreement with previous work. The release of positive charge during the growth of snow is suggested by the results of Kramer (1949) and Kumm (1951), and the release of negative charge during melting by the experiments of Dinger and Gunn (1946), although their association of the effect with the presence of air entrapped in the ice particles makes it uncertain whether or not their results are applicable to snow. Nakaya and Terada (1934) measuring the sign of the charges on individual snowflakes found that if the flakes were dry, they were predominantly negative, whereas when they had water droplets attached, they were often positive. Measurements at the ground in quiet conditions indicate excess negative

charge on snow (Simpson 1942, Chalmers and Little 1947) and excess positive charge on rain (Simpson 1949). Kuettner (1950) and Reynolds (1954) both found that, in measurements from mountains, the precipitation tended to be of opposite sign to the potential gradient. During point discharge, results at the ground will be complicated both by the presence of space charge, and by the collection of charge by the rain or snow. This is well demonstrated by Simpson's (1949) results for the total rain current, and by Hutchinson and Chalmers (1951) for the charge on individual drops. However, the present measurements suggest that the charges on precipitation are fundamental in determining the growth of fields, and that the charging of precipitation by ion capture is a secondary process which will, like the primary process, again give an inverse relation between field and precipitation charge. This is the conclusion reached by Chalmers (1956) from his total current observations.

The release of negative charge on the melting of snow gives a simple explanation of the negative charge which was sometimes found near the ground during steady rain by Chauveau (1900) and Thomson (1860 a, b). Similarly, we can explain the remark of Whitlock and Chalmers (1956) that with showers of not very great activity the negative charges associated with the cloud often appear to lag behind it. The charges will, in fact, often be released below the cloud base.

Finally, the simplicity of the field measurements on the glacier emphasizes the necessity for care in interpreting field measurements made under normal conditions where the presence of space charge and the distance of the cloud may obscure otherwise simple results.

Acknowledgment

I should like to thank the National Research Council of Canada, which financed the expedition on which these measurements were made.

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Correspondence

551.501.9 : 551.582.2 (42)

Effects of the change of site on the climatological observations at Camden Square, London

By J. Glasspoole

At a Special General Meeting of the Royal Meteorological Society held in the Society's rooms on 19 December 1957, it was agreed that 62 Camden Square should be sold because the maintenance of the property had become an increasing drain on the available funds. Satisfactory arrangements had been made with the St. Pancras Borough Council for the climatological station to be transferred some 150 yards to the Gardens in Camden Square. The new station became operative in June 1957, the full observations at 62 Camden Square continuing until 23 March 1958, so that overlapping records are available for about 10 months. While a longer overlap would have been advantageous, there is already sufficient information to indicate the effect of the change. A better comparison will be possible in 5 or 10 years' time, by using the records at other stations in the London area.

In 1858 the late G. J. Symons, F.R.S., set up a climatological station at his home at 136 Camden Road, a few hundred yards distant from 62 Camden Square, to which he moved 10 years later. When the Meteorological Office took over the station in 1919 consideration was given to its continuation in addition to those at Kew and Greenwich. It was clear that an adequate representation of the variations of rainfall and temperature over London required at least these three stations. The three stations Camden Square, Kew and Greenwich were at altitudes of 111 ft, 18 ft and 149 ft respectively; the record at Camden Square representing the higher, northern suburbs, Kew the valley near the river to the west of London in rural surroundings, and Greenwich an isolated hill selected for astronomical purposes to the east of London. The station at Greenwich terminated in July 1952 when the Observatory moved to Herstmonceux, but the Director of the National Maritime Museum continued the record at a site nearer the river at an altitude of 24 ft. The site at Camden Square was that of a small, walled, suburban garden surrounded by rows of houses. The site in the Gardens is about 10 ft higher on ground sloping up slightly to the north-east. It is more open than that at 62 Camden Square and is mainly grass. It is dominated by six trees about 60 ft high, the open space being about 40 × 33 yards, and the trees subtend angles of about 40°-45° at the site.

Rainfall

No comparable rainfall records are available for the Gardens from 8 February to 10 March 1958, because the raingauge was stolen, but with the exclusion of those 31 days the gauge in the Gardens recorded 19·34 in. from 1 June 1957 to 22 April 1958 while that at 62 Camden Square caught 19·73 in., a difference of 0·39 in. The 1916-1950 average at 62 Camden Square is 25·15 in. and simple proportion would lead to an average appropriate to the site in the Gardens of 24·64 in. or 2 per cent less than that at the old site. There were nine days on which the amounts at the two stations differed by 0·02 in. or more, and only one with a difference exceeding 0·03 in., viz. 12 August when 62 Camden Square gave 1·27 in. and the Gardens 1·18 in. On one of these nine days the Gardens recorded more than 62 Camden Square. There is no apparent reason why there should be any real difference in the long-term rainfall at the two sites, and it seems, at present, that the existing rainfall averages for 62 Camden Square should be retained without applying any correction. The slightly smaller rainfall at the Garden site is fairly uniform throughout the year, no seasonal change in the difference being apparent. The differences were larger for the less windy months, suggesting a chance variation of rainfall rather than any real difference of exposure. The differences in the monthly totals are given at the end of Table 1.

Temperature

The dry- and wet-bulb temperatures at 09 GMT were similar but showed some persistent differences. On the average the dry-bulb readings were 0·2°F higher in the Gardens, while the wet-bulb readings were lower in the summer months, June to September, but higher in the winter months, October to March, giving for the whole period similar mean values. The relative humidity was generally somewhat lower at the Gardens, on the average by about 1 per cent. This is

accounted for by the higher dry-bulb temperatures in the Gardens. Presumably, at the more open site in the Gardens, the diurnal increase of temperature becomes more apparent there at 09 GMT than at 62 Camden Square, where the air is more stagnant and less open to the sun, being restricted by the surrounding buildings and garden walls. It was not possible to take the readings at the two stations simultaneously at 09 GMT, but the observer was asked to visit one station first on alternate days. The differences in the dry-bulb readings amounted to 2°F or more on six days, on each of which the Gardens gave the higher readings. These were four days in June 1957, 31 July 1957 and 3 March 1958, and all were quiet rather than windy.

Maximum and minimum temperatures

The monthly mean maximum temperature was consistently lower at the Gardens, on the average by 1·3°F. For the months June to October and March the mean difference was 1·9°F and in the winter months November to February only 0·7°F. The differences were 3°F or more on 69 days, and these occurred in all months, apart from December and January, and with temperatures in the Gardens as low as 35°F. On 25 days the differences were 4°F or more, on 4 days 5°F or more and on one day 6°F. The large differences in the maximum temperatures occurred on quiet rather than windy days.

The monthly mean minimum temperature was consistently higher in the Gardens, but on the average only by 0·15°F. The Gardens gave the higher reading by 2°F or more on 13 days and 3°F on three days, while the readings were lower by 2°F on one or two days.

The more uniform temperatures in the Gardens are again due presumably to the more open site, the more sheltered site in the walled garden at 62 Camden Square giving higher maxima, especially in the summer, and slightly lower minima. It would be unwise to correct existing averages by this short overlap, but averages for the new station should be recomputed when records are available for 3 years and preferably longer.

Grass minimum readings

Readings were higher in the Gardens by 3°F in the summer months June to August, but by less than 1°F during the autumn. Unfortunately the grass minimum thermometer was broken in December and it was considered unwise to replace it until the risk of further damage could be decreased by erecting barbed wire on the fence round the enclosure.

TABLE 1. DIFFERENCES BETWEEN MEAN MONTHLY VALUES AT 62 CAMDEN SQUARE AND THE GARDENS IN THE SQUARE

Period June 1957 to March 1958 (difference Gardens minus 62 Camden Square)

Time (GMT) 1957	Dry-bulb temp. 09 °F	Wet-bulb temp. 09 °F	R.H. 09 per cent	Max. temp. 09-09 °F	Min. temp. 09-09 °F	Grass min. 09 °F	Earth temp. 1 ft 09 °F	Earth temp. 4 ft 09 °F	Rainfall in.
June	+ 0·6	− 0·4	− 3·7	− 2·3	+ 0·2	+ 3·4	+ 5·4	+ 5·8	+ 0·02
July	+ 0·1	− 0·2	− 1·4	− 1·7	− 0·3	+ 3·2	+ 4·9	+ 8·3	− 0·03
August	+ 0·1	0·0	− 0·2	− 1·7	+ 0·3	+ 2·7	+ 5·0	+ 8·3	− 0·13
September	+ 0·1	− 0·1	− 1·0	− 1·8	+ 0·4	+ 0·4	+ 3·2	+ 6·1	− 0·01
October	+ 0·6	+ 0·3	− 1·6	− 2·1	+ 0·4	+ 1·1	+ 1·5	+ 4·0	− 0·07
November	+ 0·2	+ 0·1	− 1·8	− 0·9	+ 0·1	+ 0·9	0·0	+ 1·8	− 0·07
December	+ 0·3	+ 0·3	+ 0·3	− 0·7	+ 0·2	–	− 1·1	− 0·5	0·00
1958									
January	+ 0·0	+ 0·1	+ 0·4	− 0·4	0·0	–	− 1·4	− 1·0	− 0·07
February	+ 0·2	+ 0·0	− 1·3	− 0·9	0·0	–	− 0·7	− 0·9	− 0·01*
March	+ 0·1	+ 0·2	+ 0·4	− 1·9	+ 0·2	–	− 0·3	–	− 0·02*
Mean	+ 0·2	+ 0·0	− 0·9	− 1·3	+ 0·1	–	+ 1·7	–	–

* incomplete months

Earth temperatures at 1 ft and 4 ft

The comparison in Table 1 shows that the Gardens gave much higher temperatures at 1 ft and 4 ft from June to November, but lower values in the remaining months. There are many cases on record where earth thermometers on nearby sites have given appreciable differences, owing to differences in the type of soil, or height of the water table. The site in the Gardens gives a further complication in that it is above the side of a tunnel along which steam trains pass to and from St. Pancras. Presumably the ground there is cooled by the air in the tunnel in the winter. The comparison is far too short to indicate the controlling factors, and readings covering a long period should prove of considerable interest when compared with similar readings made at other stations in London.

We have an opportunity of defining more closely the differences in the weather experienced in the suburban garden at Camden Square and in the small public gardens nearby, and extending our knowledge of micro-climatology. Much will depend on maintaining the interest of the observers in continuing careful observations. This could be attained by Fellows making use of records. It would be helpful if a Fellow living in the district would take a personal interest in the station and visit it from time to time by arrangement with the officers.

87 Mostyn Road,
Merton Park,
London, S.W.19.
18 August 1958.

551.501.81 : 551.509.324 : 551.515.83

The movement of precipitation belts as observed by radar

By W. G. HARPER and J. G. D. BEIMERS

(Read 15 October 1958. See *Q.J.*, **84**, p. 242)

Dr. J. M. STAGG : In presenting this paper Mr. Harper seems to me to have shifted the emphasis from belts of rain to fronts, and it is rather important to know whether this is intentional. The paper, as printed, deals with the movements of rain belts which are indeed all that the authors could legitimately consider because they did not make any synoptic study of the fronts with which the rain might have been associated. But in his presentation, Mr. Harper has related the findings of the paper to his idea of the kinematics of a cold front. This could lead to confusion.

I would like to be assured that Mr. Harper does not mean us to identify what he deduced from his PPI tube with the progress of fronts. The apparent movement of the rain could be quite different from the movement of the front to which the rain might nevertheless have been related.

Prof. P. A. SHEPPARD (*President*) : Is there evidence for, and what are the authors' views on the hypothesis that the advection of precipitation belts by the 700 mb wind normal to the belts is due to the precipitation-forming mechanism becoming effective at this, rather than at a higher or lower level ? I do not, of course, suggest that the kinematic analysis at the end of the paper is irrelevant, but it is not a complete explanation.

Dr. R. S. SCORER : Could Mr. Harper make it clear to what extent it was possible to see whether a rain cell was growing sideways (or in some other direction) relative to the wind. This can easily be seen visually, and is very important when the winds are light, as for instance when thunderstorms begin over hills and move over valleys in the evening : but how much is known from radar observations in moderate winds ?

Mr. J. S. SAWYER : What proportion of the cold fronts and occlusions which pass over SE England give well-defined rain belts suitable for inclusion in a study such as this ?

Dr. R. C. SUTCLIFFE, F.R.S. : The result of the paper is, of course, interesting and useful but we should be careful not to read too much into it. The cases studied are of fronts of cold-front type which are associated with a rain belt moving coherently : these are the synoptically simple cases. Fronts with erratic outbreaks of instability rain, or katafronts without rain must not be presumed, on the evidence, to move according to the same rule.

Dr. R. C. RAINEY (*communicated*) : In this paper the potentially objective evidence of radar is classified on the criteria of current synoptic analysis, in which one imagines that there must inevitably be some subjective element. Would the authors care to comment on any possibilities there might be in a converse approach, such as a comparison of the timing and type of all passing fronts, as indicated on the synoptic charts over a given period, with the timing and nature of all radar displays seen during the same period ?

Wg. Cdr. R. M. POULTER (*communicated*) : Having used the 10,000 ft wind as the key to many difficult weather situations, it is gratifying to see the high correlation shown at 700 mb in Fig. 2 of the authors' paper. It has been my belief for some time that frontal rain is made up of 'packets' of rain or showers moving *along* the front with the thermal wind; accordingly, to forecast details of rain intensity at a given place, such as Wimbledon, for the ensuing few hours, it is not enough to measure the movement of the front, as along AB in Fig. 1, but one must base forecasts of precipitation intensity on movements along CD. Actually, in Fig. 1 one can draw four other lines of showers sensibly parallel to CD and from that map can read off a velocity of 33 mi hr^{-1} from 288°. Next, reference to the *Daily Aerological Bulletin* shows that the wind at 10,000 ft over Larkhill at 1100 GMT on 4 July 1951, was 30 kt from 290° !

Mr. W. G. HARPER (*in reply*) : Dr. Stagg rightly stresses that our measurements were made on the precipitation and not on synoptically-defined fronts. I had no intention of moving this emphasis. In looking for a reason for the high correlation with the wind component at 700 mb, however, it seemed to us to be legitimate to consider whether the derived pattern of mean wind components would apply to a simple frontal model, and we included this in the paper. Dr. Stagg's suggestion that the movement of the rain may be quite different from the movement of an associated front could be checked only by a much more extended study than ours. It would probably require

radar evidence from more than one station if belt movements over distances on the scale of the synoptic measurements are to be studied. It is interesting that Austin and Blackmer (*Bull. Amer. Met. Soc.*, **37**, 1956, p. 447) have suggested that prefrontal bands may be formed in this way, by the movement of precipitation through and ahead of the front, but there is, as yet, no reliable evidence to confirm this.

In reply to Prof. Sheppard, the records we used in this study give no indication of the levels at which the precipitation-forming mechanism was initiated and maintained. We hoped that our subdivision of material on the basis of height of echo tops would throw some light on this, thinking that cases with high echo tops might correlate better at a higher level. That no significant difference in correlation level was found would seem to be indirect support for Prof. Sheppard's hypothesis. I feel somewhat diffident about referring to the frontal model after Dr. Stagg's comments, but I think it is interesting that warm air would be overtaken by this hypothetical front at all levels below 700 mb, implying that the maximum vertical transport of air would then occur close to the 700 mb level. Could this be a contributory reason for a precipitation-forming mechanism becoming effective at that level ?

In reply to Dr. Scorer, we could rarely distinguish the individual cells in the belts of precipitation studied in this paper, but in thunderstorm situations cell growth can usually be clearly seen on radar. To find the effect of topographical features in moderate winds requires very detailed study, because the precipitation only becomes detectable by centimetric radar when it has reached raindrop size, by which time a substantial horizontal movement can have taken place. A study of the distribution and movement of showers was attempted at East Hill by Mr. R. F. Jones (unpublished) some years ago, which included an appreciable number of showers on days with moderate winds. Their movement was projected backwards, but no obvious relationship to orographic features was found. We have recently undertaken a further study using thunderstorm material, but it is too early to give any result from this.

In reply to Mr. Sawyer, I would think that in about a half of the occurrences of frontal precipitation (using the rather loose definition of ' association ' with a front that we adopted in our paper) a reasonably well-defined edge of precipitation is seen in some part of the display which is suitable for measurement. This is based on experience with a long-range PPI operating at 10 cm wavelength, for which attenuation by intervening rain is negligible. The rather small number of belts we have included in this paper, as compared with the number of cold fronts and occlusions which must have passed over southeast England in the period we used, results not from any process of exclusion, but from the fact that on average our radar was not in operation for more than 6 hours out of the 24. I must agree with Dr. Sutcliffe, however, that we were probably dealing with the synoptically-simple cases.

In answer to Dr. Rainey, I suggest that the greatest benefit should accrue from co-ordination of synoptic and radar information by the forecaster, if the latter could be made available to him, since, within the range of the radar, one can appreciate the extent and type of precipitation far better than from the synoptic network.

Finally, I agree with Wg. Cdr. Poulter that in making a short-period forecast for a point one must study the individual cell movements, if necessary by reducing the gain of the receiver to identify them. The variability of rainfall along frontal precipitation belts is a pronounced feature of the radar record.

551.515.23

Modification of the structure of a tropical cyclone on traversing a high plateau

By N. P. SELLICK

(Read 15 October 1958. Presented by Dr. A. G. Forsdyke. See *Q.J.*, **84**, p. 259)

Dr. R. S. SCORER : The heat released by the condensation of the rain was of the order of 10^{26} ergs a day, or one H-bomb every 15 minutes. This estimate of the energy of the storm does not include the energy of condensation of all the ice cloud which flows out at the top of the storm.

The trouble with the calculation of the dissipation of kinetic energy is that there are no upper-wind measurements in the region of outflow so that no estimate can be made of the dissipation due to the exchange of momentum between the lower and upper halves of the storm. I would guess that dissipation on this account would far exceed that due to friction at the ground, and the author's estimate of c could quite easily be out by a factor of 10.

Dr. R. C. Rainey (*partly communicated*) : I should like to enquire whether differences of air-mass may have played any part in this very informative story. The conjunction of surface currents from NW and SE, a common feature of rainy spells in Rhodesia, as the author mentions, and also in the Transvaal and the Orange Free State (Rainey 1945; *S.A.A.F. Met. Sect. Tech. Note* No. 45), often appears to be associated with a difference in humidity, with dew-points higher in the NW 'Congo air' than in the south-easterlies from the Indian Ocean (Southern Rhodesia, 1953, *WMO I-RAI*, pp. 166-172). Such a difference, if present on this occasion, might perhaps have contributed to the observed distribution of rainfall, largely to the north of the track, since any participating air-masses might be expected to retain their identity longer in these circumstances than at the much greater angular velocities of a tropical cyclone. Kreft (*Notos*, **2**, p. 149, 1953) indicates that dew-points were generally high over this period, but the intrusion of moister air during the passage of the depression, shown by the radiosonde data for Salisbury (south of the track and apparently in surface south-easterlies and easterlies most of the time), suggests that appreciable differences in humidity may still have been concerned. In view of the particular interest of the quantitative estimate of the convergence involved, it would also seem desirable to know to what extent the ascent of 18 February may be taken to have been representative, not just of conditions over Salisbury during this period, but of the system as a whole.

Another example of the modification of a tropical cyclone over land in the same area and at the same time of year, is illustrated by the barograph charts shown, which relate to the passage of another cyclone, westwards across Madagascar at 18°-19°S (cf. author's Fig. 3) during 12-13 February 1943. At Tamatave, on the east coast, the anemometer was reported to have failed at a windspeed of some 110 kt, and H.M. Trawler *Pladda* (545 tons) was 'driven high ashore'; at Ambohitsilaozana, 100 km inland from Tamatave, a wind force of 11 (60 kt) was reported, and at Tsiroanomandidy, 360 km inland, one of force 10 (50 kt). The rains continued during the passage of the cyclone across Madagascar, and one of the heaviest recorded falls was in fact on the west coast (Besalampy) on 13 February 1943.

Exceptional Indian Ocean cyclones can at times bring heavy rains even into the interior of Arabia; a storm of this nature, which had been detected near the Maldive Islands on 19 October 1948, reached the Arabian coast near Salala on 25 October, and was followed by heavy rains between 26 and 31 October over a wide belt of the Rub-al-Khali sands in the interior across to the Persian Gulf.

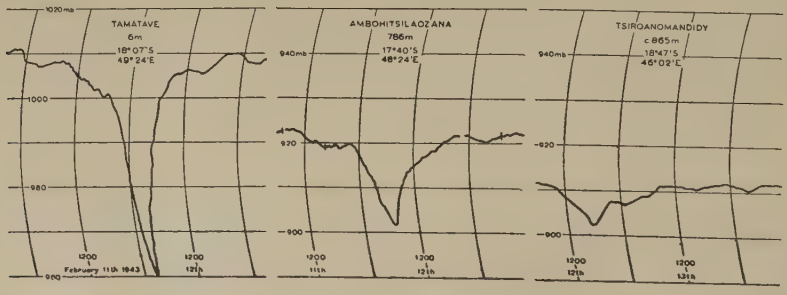

Figure 1. Barograph records for coastal and inland stations during the passage of a cyclone across Madagascar (with acknowledgments to the Service Météorologique de Madagascar).

Mr. J. S. Sawyer : This paper is an interesting demonstration that this tropical cyclone on passing overland continued to derive its energy from the release of latent heat in the large cloud mass. The energy of the storm overland was not merely derived from the kinetic energy developed over the sea; the role of the original oceanic cyclone was to organize the large cloud mass and initial circulation which was subsequently self-maintaining, but could not have readily developed overland.

Dr. H. L. Penman : It was stated that the convergence area had twice the diameter of the rain area. Is this as surprising as Dr. Forsdyke made it sound ? To get the intense rain measured near the centre I presume that there must be a rainfall area outside, several times as big, over which the converging air picks up its water vapour. Is this presumption unreasonable, or contrary to experience ?

Dr. R. C. Sutcliffe, F.R.S. : The paper may be interpreted as a verification of an estimate of the frictional dissipation of energy. It would be interesting on theoretical grounds to know whether the very rapid decrease of circulation, as the cyclone moved from the sea to land, could be explained as the effect of the increase of friction – but the effect of land elevation is a complication.

Mr. C. E. Wallington : The analysis by Mr. Sellick presents some interesting points for discussion, but why limit an explanatory hypothesis to the creation of vorticity by convergence and its dissipation by friction ? It would be just as interesting and relevant to conjecture how the upper-wind structure could have fitted into the picture and whether or not the influx of moisture could be accounted for by studying trajectories rather than instantaneous streamlines.

Mr. N. P. Sellick (*in reply, communicated*) : I wish to thank Dr. Forsdyke for presenting this paper in my absence.

Dr. Scorer discusses the estimate of the rate of dissipation of kinetic energy. In the numerous assumptions made, I endeavoured to find the minimum flow required to produce the observed rainfall. The adopted rate-of-loss of kinetic energy was based on assessments made by others and the fact that it satisfied the calculations was a surprise to me. The use of a less efficient flow pattern could readily accommodate higher estimates of the rate-of-loss of kinetic energy but I doubt whether it would stretch to ten times.

I would draw Dr. Rainey's attention to the velocities in Table 4. At the shorter radii one complete revolution takes about 1·8 days and surfaces of air mass discontinuity would be wrapped up like a Swiss roll. I have seen something like this in radar pictures of oceanic cyclones. It is regretted that upper data other than those for Salisbury are not available. On 15 February when the cyclone was on the coast, and after 20 February, when it was receding, there was dry air aloft over Salisbury. The change appeared to be progressive as is usual here, not sharp.

I do not follow Mr. Sawyer. I took it for granted that the copious rainfall was evidence of the continuing release of energy and sought reasons why this energy was not reflected in increasing pressure gradients and winds at the surface.

Dr. Penman raises a point which has interested me very deeply and a full reply would be unduly long. Great quantities of desiccated air are produced over an area of intense rain. The distance to which this air can be removed is limited by the velocities developed by motion down the pressure gradient and a complete discussion of the motion associated with the maintenance of a rain area on a synoptic scale must include the disposal of this air. This involves computations of accelerations, descent, radiation losses, additions of moisture, turbulence and friction losses and other factors. I have found that the appearance of a jet stream is a necessary consequence of the disposal of the air used in maintaining a belt of rain. It follows that air approaching a rain area from some distance must have suffered descent and it must suffer convergence and the addition of moisture from the ground before it produces rain.

In reply to Dr. Sutcliffe, the paper originated in an argument as to whether the cyclone observed over the ocean and the disturbance over the land were the same system. I attempted to examine the changes as the cyclone moved from the sea to land but was defeated by lack of data and turned my attention to the later phase where data were less inadequate.

I do not follow Mr. Wallington. The essence of the problem is that whereas argument from the rainfall would lead, in coordinates moving with the system, to trajectories differing from the instantaneous streamlines, i.e., development, observation showed that there was no appreciable change in the pressure field and hence, by inference, in the wind field. Thus the streamlines and the trajectories were sensibly identical. The equations in para. 12 are essentially dynamic but the accelerations due to the inertial effects are opposed by those due to friction and turbulence.

Dr. A. G. Forsdyke (*communicated*) : Mr. Sellick in his reply to the discussion has replied to the points raised more adequately than I did at the meeting and I have nothing to add to what he says. I would however reiterate that this paper is a good example of how the difficulties arising in orthodox synoptic analysis in low latitudes and over plateau territory where observations are sparse in space and time may be met by special methods. In this case it is done by using station-level pressure anomalies and time sequences of observations at a few stations to build up a single picture of a disturbance assumed to be moving steadily without change of structure.

Proceedings at meetings of the Society

Ordinary Meeting, 15 October 1958

The following candidates were declared to have been elected by the Council during the summer recess :

Fellows

Bashorun, Buliaminu Babajide, Meteorological Office, Ikeja, Nigeria, B.W. Africa.
Chapas, Leslie Constantine, B.Sc., c/o West African Institute for Oil Palm Research, Nr. Benin City, Nigeria, B. W. Africa.
Draper, Lawrence, M.Sc., National Institute of Oceanography, Wormley, Godalming, Surrey.
Eyo, Ita, Meteorological Service, Minna, Nigeria, B.W. Africa.
George, David John, Holcombe Farm, Painswick, Nr. Stroud, Glos.
Hardy, Joan Evelyn, B.A., 6 Brunton Terrace, Sunderland, Co. Durham (*transfer from Student Membership*).
Hearn, Jacqueline Margaret, 7 Penare Road, Penzance, Cornwall.
Imber, Robert George, 65 Ankerdine Crescent, Shooter's Hill, London, S.E.18.
Jagger, Harold, 185 Farnborough Road, Farnborough, Hants.
Lord, Maurice Kenneth, 89 Skellingthorpe Road, Lincoln.
O'Byrne, John W., M.Sc., A.Inst.P., 17 Lanahrone Avenue, Corbally, Limerick, Ireland.
Paton, Eleanor Anne, Greenacre, Croft Avenue, Dorking, Surrey.
Reed, Richard Raddon, B.Sc., c/o District Commissioner, Butha Buthe, Basutoland.
Roberts, Cedric Geoffrey, 85 Hagley Road, Hayley Green, Halesowen, Birmingham, Warwicks.
Wake, Peter Gerald, B.A., 9 Spring Road, Ipswich, Suffolk.
Young, Capt. John Reginald Cass, 81 The Drive, Isleworth, Middx.

Foreign Members

Franceschini, Asst. Prof. Guy Arthur, M.S., 112 Fleetwood, Bryan, Texas, U.S.A.
Staley, Asst. Prof. Dean Oden, M.A., Ph.D., 1008 Gilson Street, Madison 5, Wisconsin, U.S.A.
Vederman, Joseph, M.S., 477 Madison Street N.E., Washington 11, D.C., U.S.A.

Student Members

Buckle, Mrs. Beryl Kathleen, 58 Scarsdale Villas, London, W.8.
Collar, Joan Terry, B.Sc., 97 Trinity Road, London, S.W. 19.
Hughes, Marilyn Elisabeth, 369 Gower Road, Sketty, Swansea, Glamorgan.
McHardy, Frank John, Craigendarroch, St. Michaels, Nr. Leuchars, Fife, Scotland.
Smith, John Charles, Queen Anne Inn, Queen Anne Road, Maidstone, Kent.
Stewart, John Brampton, 8 Cope Place, Kensington, London, W.8.
Wardhaugh, Keith Grant, B.Sc., 54 Hollywell Road, North Shields, Northumberland.
Yardley, Beryl, 45 Teddington Park Road, Teddington, Middx.

Canadian Branch

Fellows

England, Mervyn Clarence, M.A., 32 Fairview Blvd., Toronto 6, Ontario.
Machattie, Leslie B., M.A., Forest Research Division, Forestry Branch, Motor Bldg., Ottawa, Ontario (*transfer from Associate*).

Associates

Eddy, Fred W., M.A., Main Meteorological Office, Trans-Atlantic Administration Bldg., Montreal Airport, Quebec.
Lemire, Francois Jacques, B.Sc., M.A., Meteorological Office, Department of Transport, Goose Bay Airport, Labrador (*transfer from Fellowship*).
Lortie, Joseph L. G., 229 Wright Avenue, Toronto.
Parent, Leonard E., M.A., District Aviation Forecast Office, Whitehorse, Yukon.
Wilson, Howard James, B.A., Aviation Forecast Office, Box 71, R.C.A.F. Station, Gimli, Manitoba.

The following candidates were then declared to have been elected that day by the Council :

FELLOWS

ASHFIELD, Kenneth Thornby, 52 Walpole Road, Old Windsor, Berks.
BARFIELD, Edward Charles George, LL.B., B.A., B.Sc., 20 Dingle Hill, Colwyn Bay, Denbighshire, N. Wales.
ROSSI, Desmond Eric, 42 Ravenswood Avenue, Tolworth, Surbiton, Surrey.
SHIRVELL, Jack Baden, 20 Pembury Road, Westcliffe on Sea, Essex.

The following papers were read and discussed :

1. 'Movement of precipitation belts as observed by radar' by W. G. Harper and J. G. D. Beimers (*Q.J.*, **84**, p. 361).
2. 'Modification of the structure of a tropical cyclone on traversing a high plateau' by N. P. Sellick (*Q.J.*, **84**, p. 259).

Evening Discourse, 5 November 1958

The first of a series of Evening Discourses took place in the Society's Rooms; Dr. B. J. Mason spoke on 'Recent developments in the physics of rain and rainmaking.'

A full account of this meeting appeared in December 1958 issue of *Weather*.

Ordinary Meeting, 19 November 1958

The following candidates were declared to have been elected that day by the Council :

FELLOWS

ALDIS, Robert Harold, P.O. Box 87, Accra, Ghana.
BRADBURY, Thomas Arthur Michael, Bremhill Vicarage, Calne, Wilts.
DWEMOH, Edward Richard, Department of Civil Aviation, P.O. Box 87, Accra, Ghana.
HALL, William Shield, 18 Onslow Terrace, Langley Moor, Co. Durham.
HEEPS, Thomas Noel, B.A., 80 Meadowbank Road, Remuera, Auckland, S.E.2., New Zealand.
LANDBECK, Roger Charles, 35 Grosvenor Gardens, Woodford Green, Essex.
MILLER, Derek Edward, B.A., 22 Marlowe Close, Banbury, Oxon.
MONRO, Capt. James, Alsterlea, Eghams Wood Road, Beaconsfield, Bucks.
OSBORNE, R., 27 Eastbourne Street, Monks Road, Lincoln (*reinstatement; elected 1943, resigned 1951*).
ROWE, Lt. Cdr. James Edward, R.N., B.A., H.M.S. *Ganges*, Ipswich, Suffolk.
SHOENBERG, Mark Ely, B.A., 20 Ashfield Road, East Acton, London, W.3.
WHEATLEY, Collin Henry, Meteorological Office, Department of the Interior, Madang, Territory of Papua and New Guinea.

FOREIGN MEMBERS

FUNK, Josef Peter, D. Phil., C.S.I.R.O., Division of Meteorological Physics, Station Street, Aspendale, S.13., Victoria, Australia.
PIERRARD, John Martin, B.S., M.S., P.O. Box 6905, College Station, Texas, U.S.A.

STUDENT MEMBERS

ASSADI, Adel, Department of Meteorology, Imperial College, Exhibition Road, London, S.W.7.
DUNMORE, Richard, 41 Prince of Wales Road, Coventry, Warwicks.
EL-SAIYED, Hassan, Department of Meteorology, Imperial College, Exhibition Road, London, S.W.7.
ROGERS, Peter Richard, 111 Hazelbank Road, Catford, London, S.E.6.
SMITH, Hilary, 44 Nottingham Place, London, W.1.
WOOD, Joan, 33 Huddersfield Road, Diggle, Dobcross, Nr. Oldham, Lancs.

The meeting took the form of a Discussion on 'Atmospheric chemistry.' The opening speaker was Mr. B. C. V. Oddie who gave a broad survey of the chemistry of the lower atmosphere. Prof. H. Egnér (Institute of General and Inorganic Chemistry, Uppsala, Sweden) then discussed the gaseous equilibrium between the atmosphere and the oceans, biosphere and soil. In the evening session Dr. Junge (U.S. Air Force Research Center, Cambridge, Massachusetts) discussed the origins, distribution and reactions of the aerosols normally found in the atmosphere.

Special General Meeting, 17 December 1958

Professor P. A. Sheppard, President, in the Chair, and about 65 Fellows present. The President said that, following discussions in Council that afternoon, he would recommend to the Meeting that the Business arranged should be postponed to a Special General Meeting or Meetings to be called as early as practicable.

The Meeting had been called to approve certain amendments to By-laws which were necessary to implement recommendations of Council. However, after the Notice calling the Meeting had been issued, a letter was received from the Canadian Branch expressing some anxiety about the consequences of abolishing the Associate grade of membership. The President explained that these matters had been discussed earlier in correspondence with the Canadian Branch but, in view of their latest letter and of their large Associate membership, Council considered that further consultation with the Canadian Branch should take place before the By-laws were amended.

The Meeting then agreed to the postponement of the Business.

Ordinary Meeting 17 December 1958

The following candidates were declared to have been elected that day by the Council.

FELLOWS

ALLENBY, Reginald Thornton, B.Sc., 25 Hastings Drive, Flixton, Urmston, Manchester, Lancs.
BACON, Patrick John, B.A., 17 Orchard Avenue, Gravesend, Kent.
BAYLISS, John Michael, ' Halcyon,' Aiskew, Bedale, Yorks.
CADE, Cecil Maxwell, A.M.Brit.I.R.E., A.I.E.E., M.I.N., M.R.I., ' Genova,' 42 Latton Green, Harlow, Essex.
PERRIN, Brian, 56 Ferndene Road, Herne Hill, London, S.E.24 (*transfer from Student Membership*).

STUDENT MEMBERS

BAKER, Michael John, 16 Brashfield Road, Bicester, Oxon.
BYE, John Arthur Tristram, 82 Brunswick Road, Ealing, London, W.5.
TAYLOR, Sheila Mary, 36 Cheviot Avenue, Oldham, Lancs.

FOREIGN MEMBER

WARTENA, Lambertus, Diepenbrocklaan, 26, Arnhem, Holland.

The following papers were read and discussed :

1. ' Some observations on flying locusts and atmospheric turbulence in Eastern Africa ' by R. C. Rainey (*Q.J.*, **84**, p. 334).
2. ' Vertical distribution of aphids in the air and the temperature lapse rate ' by C. G. Johnson (*Q.J.*, **83**, p. 194), presented by Dr. H. L. Penman.

Popular Lectures

20 November 1958. Mr. P. J. Meade (Secretary) took the Chair at the first of the Society's series of Popular Lectures for 1957-58 which was given by Dr. E. T. Pearce, whose subject was ' Thunderstorms.' The Lecture was held for the benefit of sixth-form pupils drawn from about fifty schools in the London County Council area and was held in the Conference Hall, County Hall, by invitation of the Chief Education Officer of the L.C.C.

The Lecture, which was illustrated by slides, was attended by about 450 pupils, and by the high standard of questions put to Dr. Pierce during discussion, proved once again the popularity of this series of Lectures.

21 November 1958. Dr. G. D. Robinson took the Chair when Dr. E. T. Pierce repeated his Popular Lecture, which was again held in the Conference Hall of the London County Council, and which was attended by about 470 pupils from schools in the Home Counties, including Buckinghamshire, Essex, Hertfordshire and Surrey. This Lecture was held by arrangement with the Education Officers of the Councils from which the pupils were drawn.

The total attendance at the two Lectures at the Conference Hall was about 920.

Reviews

Geophysics and the I.G.Y. : Geophysical Monograph No. 2, American Geophysical Union; (Proceedings of a symposium at the opening of the International Geophysical Year). Baltimore. Eds. H. Odishaw and S. Rottenberg. (The Lord Baltimore Press), 1958. Pp. v, 210. Figs.; Tables. $8.

It was a happy idea of the United States committee for the international geophysical year to hold a symposium on the eve of I.G.Y. and to publish the symposium papers as a Geophysical Monograph. The intention of the committee obviously was that each contributor should treat his subject fairly broadly, even semi-popularly, and yet should leave a clear impression both of the state of his subject as at D-day for I.G.Y. and of the main problems awaiting solution. In these aims most of the contributors have succeeded admirably; some contributors, for example E. B. Roberts and D. G. Knapp, have also managed to give an instructive historical setting to the treatment of their themes.

The Monograph is in three sections : one (18 papers) on upper atmospheric physics, a second (10 papers) on the lower atmosphere and the earth itself and a third small section of 2 papers on the polar regions. The second group of papers is the more purely meteorological of the three – if there is such a subject as pure meteorology – and even in it there is clear advantage in having oceanography and meteorology well mixed up. R. R. Revelle and W. H. Munk contribute two studies of the earth's heat and water budget, and, close to a paper on synoptic meteorology by M. J. Rubin, there is a synoptic study on oceanography by C. O'D. Iselin. There is also a paper by S. Fritz on special meteorological studies; it deals with radiation and the composition of the atmosphere. Along with other interesting topics these papers discuss the positive and negative feedback effects that an increase of solar radiation can introduce into the complex mechanism which controls air temperature, the possible effect on the melting of glacier ice (and hence on ocean levels) of the present rapid consumption of the earth's store of fossil fuels, and the providential arrival of electronic computers just when they will be needed to cope with the vast amount of I.G.Y. data.

Although the papers in section 2 are all good, the broadminded meteorologist – and is there any other kind ? – may find greater interest in the first section on the physics of the upper atmosphere. Here is a very readable collection of papers, presenting up-to-date views and problems in solar-terrestrial relations, the structure of the high atmosphere, the ionosphere, geomagnetism, cosmic rays, aurora and air glow. Whistlers get two papers to themselves, and various aspects of rockets and satellites get five. Whistlers, incidentally, are those remarkable phenomena in which low-frequency radio pulses (probably originating in lightning) get caught up by a magnetic field line in the earth's magnetic field and spend their dying moments travelling back and forward through space from one end of the line to its conjugate on the other side of the earth.

It is instructive to see how all these high-atmosphere topics are interrelated and it is salutary to be reminded that many of the basic phenomena are still but poorly understood. For example, where do the charged particles come from which, injected at high speeds into the atmosphere, produce aurora and disturbance in the ionosphere and in the geomagnetic field ? Directly from the sun ? Perhaps ! Though it is not known from which regions on the sun, except in the case of solar flares. Plausible theories maintain that the particles could equally well be leaks from the hypothetical (though geomagnetically fairly well substantiated) ring current around the earth, or even that they come from regions in the outer atmosphere itself.

Like the first number of this Geophysical Monograph series, this volume is printed in a double-column page so there is much more in it than its 210 pages suggest. It has many diagrams and photographs, and apart from one or two slips, e.g., sub-polar for sub-solar on page 16, the volume has been carefully proof-read. But surely at least the title of No. 1 deserved mention somewhere in this second of a series which is likely to be a valuable addition to the literature of geophysics.

J.M.S.

Atmospheric explorations. Ed. H. G. Houghton. Cambridge, Mass. (M.I.T. Technology Press); London (Chapman and Hall), 1958. Pp. 125. 65 Figs.; 3 Tables. 52s.

This is a collection of papers presented at a symposium arranged by the American Academy of Arts and Sciences to commemorate the 250th anniversary of Benjamin Franklin's birth.

The first three of the five contributions discuss, appropriately, the electrification of clouds and the lightning discharge. They are of uneven quality.

Ross Gunn presents a clear and well-written account of his recent researches on the electrification of cloud and raindrops by ionic capture. Kuettner, in his discussion of charge generation in thunderstorms, raises some perfectly valid points but his case for wanting precipitation to be positively rather than negatively charged in order to account for the observed polarity of a thundercloud is neither clear nor convincing. In the most stimulating of these three essays Loeb points to similarities between the lightning stroke and the mechanism of the positive-streamer spark as studied in the laboratory.

The two remaining papers deal with the upper atmosphere. H. Wexler, on the basis of synoptic analysis of particular situations, discusses the vertical propagation of large-scale disturbances and the coupling between stratospheric and tropospheric systems. H. G. Booker, in his short review of radio scattering in the ionosphere, succinctly describes a number of interesting phenomena, none of which has yet been satisfactorily explained. Since the author thinks the underlying cause to be atmospheric turbulence, perhaps this is not surprising.

Perhaps, in this modern age, it is unreasonable to expect a memorial volume to consist of new and original contributions written with a little more elegance and style than we find in the normal research paper. But it does seem a pity and a little pointless to publish, in special form, re-hashed versions of material which has already appeared (sometimes more than once) elsewhere.

B.J.M.

Meteorology of the Antarctic. Ed. M. P. van Rooy. Weather Bureau, Department of Transport, Pretoria, South Africa, 1957. Pp. xi, 240; 8 Plates; 94 Figs.; 126 Tables. Detailed contents; no index. £3 10s.

Europeans who visit the temperate Southern Hemisphere cannot fail to be impressed by the relatively cool summer temperatures they encounter, taking account of their latitude. Much of New Zealand has the summer of southern England in the latitude of Portugal; while Melbourne is almost in the same latitude as Athens. Everywhere, indeed, the south wind quickly made its presence known to the early settlers. Southerly bursters, the rude sea breeze at Valparaiso, pamperos, cool changes, the Tablecloth with its accompanying phenomena at the Cape, the vigour of the S.E. trade at St. Helena, all testify to the effects of the elevated whiteness lying around the South Pole, occupying one tenth of the earth's land surface.

The energetics of the system require elucidation. Not only the snow-covered Antarctic continent, with its high albedo, is to be accounted for; around it, the spread of relatively cold surface waters far into temperate latitudes produces the formidable climate of Kerguelen in the latitude of Paris, and of South Georgia in the latitude of Manchester. Halley himself recorded freezing temperatures far out in the South Atlantic in 1699.

The first essays towards the ordering of knowledge of the Antarctic came perforce from the most distant sources. Cook, Bellingshausen, Ross, Meinardus, Simpson and Nordenskiöld were characteristic of their times. But the scientists of the southern continents are more immediately confronted with the south wind and all its effects. It is but natural that their researches should become directed towards the Antarctic. Mossman, working for Argentina, and Kidson in New Zealand began this newer tradition

It is therefore with particular cordiality that meteorologists will salute the appearance of this volume, produced by the meteorologists of the South African Weather Bureau under an editorial committee headed by Dr. Schumann. The compilation is a most welcome development of the Bureau's Southern Hemisphere Analysis Project, with its production of daily historical weather maps based on the careful assemblage and study of a very extensive range of literature, of observations and of climatic data, from stations very widely scattered both in space and time. As a result, an annotated bibliography of source-material is later to be published. The study of the literature, however, has borne fruit in this volume; a timely publication, for it is 20 years since the appearance of Meinardus's classic work on Antarctic climatology, which in its turn came almost 20 years after that of Simpson. Timely, moreover, because it forms an indispensable handy reference for the research workers whose papers will soon be forthcoming from the I.G.Y.

In eleven chapters we find surveys of the Antarctic environment, of the sources of data and of instrumentation; discussions of pressure, precipitation and evaporation; of the climate of the ocean, the islands, the coast and of Graham Land. Lastly, there are two valuable chapters summarizing what is known of the upper air, and of the synoptic meteorology; these are well up to date, and in the latter, observations at Pionerskaya and Shackleton in July 1956 are discussed; it is odd to find Halley Bay not used. Ten of the chapters are by members of the South African Weather Bureau. The eleventh, on precipitation and evaporation, is contributed by Dr. Loewe from Australia and is characteristically constructive. Dr. E. Vowinckel contributes two chapters on the oceanic and coastal climates. We are given little on the continent itself; there could be more on questions of radiation. No doubt the I.G.Y. will give rise to a fuller discussion of the high interior, towards which H. H. Lamb has already contributed lively preliminaries.

The bibliography is extensive and appears to include all the major references up to the middle of 1956. Within it, the pages on which each reference has been cited are given; this is a most welcome feature, enabling readers to pick up the views of the several authors on many points of controversial interest. European sources are evidently appreciated and are frequently cited; the general map is based on Kisack (1955) although it is pointed out that the recent Australian and U.S. maps show some differences. The work of the Falkland Islands Dependencies Meteorological Service is well recognized. Comparing the comments by Fabricius on the meteorology of Grytviken on p. 125 with those by Burdecki on p. 154, it is not entirely clear to this reviewer that Burdecki's modification of Pepper's isotherms in the region of South Georgia will be justified. There is already (1957) a new paper by Mansfield and Glassey to add to our knowledge of the meteorology of this region (*F.I.D.S. Sci. Rep.* No. 16).

Readers will find stimulating food for thought in the numerous well-produced maps and diagrams; for example the January mean cloud amount, with its interesting variations around the Weddell Sea, and the interpretation given of the July sea-level isotherms with their varying thermal gradients around the margins of the continent. Among many enlivening comments one may note the emphasis by Nagel (p. 40) on the effect of Antarctic climatic conditions on the observer; and Hofmeyr's support for the Simpson pressure waves (p. 73). Fabricius's methods of analysis of climatological data are effective, and not often met with in present-day publications; his remarks on sea temperature in the N.E. Atlantic (p. 124) appear somewhat puzzling. The statement on p. 1 that south of 50°S meteorological observations began in 1772 might be qualified with a reference to Halley in 1699; and surely Bougainville and Kerguelen must have carried instruments ? There appear to be very few slips or misspellings; among them, for example, ' Deemes ' on p. 113 should be ' Deerness.'

Printing and production are admirable ; and without doubt the work will fulfil Dr. Schumann's hope that it will provide material of value to all meteorologists engaged with the high southern latitudes. Indeed it provides good reading; and it is a volume to be recommended to all meteorological libraries, and to many more general libraries with an interest in Antarctic problems.

G.M.

British weather in maps. By J. A. Taylor and R. A. Yates. London (Macmillan), 1958. Pp. xiv, 256. 81 Figs.; Tables. 21s.

This book is a development from the First Year Course in Practical Climatology conducted by the authors at the University College of Wales, Aberystwyth. Its general plan is to introduce the fundamental properties of air masses, fronts and pressure systems, and to illustrate them by a detailed study of maps and observations published in *Daily Weather Reports* of the Meteorological Office. Each of the weather situations discussed is shown on a map of British Isles observations, accompanied by analyses covering a larger area, and by plots of one or two tephigrams. Supplementary maps of the British Isles show isopleths of surface temperature and dew point, of the difference between them, of barometric tendency and of visibility. All this information is commented upon in detail in order to build up a coherent weather picture in relation to the synoptic situation. Air-mass characteristics are associated with the history and environmental influences of the air and frequent reference is made to the classification evolved by Belasco; frontal weather is related to idealized models

On reading this book a meteorologist soon becomes aware of what he may suppose to be the difference between his and the geographer's approach to the subject. When a meteorologist sees an observation of cloud he almost sub-consciously asks himself what caused it to develop : he sees observations as the outcome of atmospheric movements. Whether the reader who takes this book as his introduction will acquire the same approach is doubtful. Description is indeed interspersed with explanation, but the treatment of atmospheric processes is hardly adequate and not always easy to follow. For example, convection is illustrated by diagrams of environment curves and adiabatic ascent curves meeting at the same surface temperature, and the reader must be puzzled as to how they can exist simultaneously. No clear picture is given of the role of land, water and cloud in the heat exchanges which modify the temperature structure of the air. No explanation of the geostrophic relationship is attempted, and the gradient wind is not mentioned. The treatment of humidity is confused, as, for instance in the statement on p. 63 that a low relative humidity was ' not surprising, for air at low temperatures has a low capacity for moisture.' Indeed, the authors are occasionally guilty of spurious reasoning, as when they claim to account for the difference of speed and slope between cold and warm fronts (p. 121) and when they interpret a steady pressure field as indicating that there is 'just one air mass' (p. 64).

It is to be hoped that the uninitiated reader will not suppose that this book describes the forecaster's approach to the task of analysing a chart. As is made clear in the preface, the emphasis is on interpretation and on consistency between observations and analysis. Nevertheless, the studies are carried out with commendable thoroughness even though the observational picture is two-dimensional apart from brief references to stability and inversions.

C.J.B.

International Journal of Air Pollution. Vol. 1. No. 1. Pergamon Press, London, 1958. Pp. 153. Subscription per volume £6 for libraries etc., £3 10s. for private individuals.

This is the first issue of a new journal of research to be devoted to studies of air pollution and to papers in allied subjects bearing upon air pollution. Contributions are expected from meteorologists, fuel engineers, physical chemists, doctors and authorities on public health, and, indeed, we find them all in this first issue.

At the present time the output of scientific papers and the number of scientific journals is so great that one might well question the need for a new research publication. The aim of the promoters of the *International Journal of Air Pollution* is to bring into one publication contributions from the widely different disciplines which contribute to the study of problems of atmospheric pollution. This is one way of broadening the interests and outlook of those who come to the study of air pollution with necessarily specialist basic training as doctors, meteorologists and so on, and to this end a number of review articles will be published. The purpose will be served at the expense of making individual papers in meteorology and other sciences somewhat less accessible to the specialist in the field whose main interest may be other than air pollution.

In the long run the test of the value of a research journal lies in whether or not it attracts contributions of permanent significance, and in this respect the journal has made a good start. The first issue contains eleven research papers covering medicine, atmospheric chemistry, meteorology and pollution measurement. They approach atmospheric pollution from widely different standpoints. There are also two review articles – one on atmospheric chemistry by B. C. Oddie, the other on the provisions of the Clean Air Act by G. Nonhebel. The first issue concludes with 15 pages of reports of meetings and book reviews. The new venture deserves success.

J.S.S.

Principles and applications of random noise theory. By Julius S. Bendat. New York (John Wiley and Sons), 1958. Pp. xxi, 431; 107 Figs.; 15 Tables. 88s.

The author deals with his subject very thoroughly and systematically. There are summaries and bibliographies for each of the 10 chapters, and a comprehensive bibliography comprising 141 items. The subjects treated include power spectra and relationships, optimum linear prediction and filtering, statistical errors in autocorrelation measurements, and optimum time-variable filters. In the whole book the only meteorological reference is a short one, to atmospheric turbulence, but the topics discussed are relevant and important, not only in micrometeorology and instrumental meteorology, but also in climatology and long-range forecasting. Meteorologists who are prepared to master the largely unfamiliar terminology will find this a valuable reference work, summarizing many fundamental researches of which the original accounts are not readily available in this country.

J.M.C.

Introduction to meteorology. By Sverre Petterssen, New York (McGraw-Hill), 1958. Pp. ix, 327, 213 Figs.; 14 Tables; 52s. 6d.

This second edition of *Introduction to Meteorology* typifies Professor Petterssen's lucid style of writing a well illustrated textbook which achieves its stated purpose – that of acquainting readers new to meteorology with the facts, features and economic significance of weather and climate.

Several sections of the book, including chapters on the composition of the atmosphere, instruments and observations, cloud types, heat and temperature changes and the equations of motion differ from their counterparts in other textbooks only in detail and in style, but under such headings as convection and precipitation, radiation and exchange of heat, showers, thunderstorms, hail and tornadoes and tropical disturbances, discussion is enhanced by the findings of recent research and by the inclusion of particularly instructive observations. Chapters on climatology are impressive for their clarity and illustrations, two of which are in colour.

Unfortunately the presentation of fronts and air masses is outdated. Although the upper structure of elementary frontal models are shown, the treatment of frontogenesis and cyclogenesis is basically two-dimensional. Furthermore, the two processes are described almost as though they are linked by fortuitous coincidence. Here the lucid style may well instil into the newcomer to meteorology an immature and inflexible idea of fronts and air masses. Of course, justice can seldom be done to a complex process in an elementary textbook, but it is disappointing to find such an able tutor attempting little more than the teaching of twenty years ago.

A chapter on local winds is rather superficial in that it describes local winds such as sea breezes, mountain and valley winds and drainage winds with only a cursory mention of their causes and effects; little is done to arouse the interest of those students who may so easily be tempted to enliven their meteorology by local studies.

Of course, these criticisms are personal opinions. Decisions on what to include and what to leave out of a textbook are best made by the author who writes for a specific class of reader. Professor Petterssen's achieved aim in this book appears to be that of introducing copious items of information (including some weather lore and the derivation of technical terms) from which the critical or tutored student can choose a selection for further study.

C.E.W.

Meteorologische Abhandlungen, der Inst. f. Met. u. Geoph. Bd. IV, Heft 1, 2 and 3. Berlin (Dietrich Reimer), 1958. Heft 1 and 2. (By I. Jacobs) 5-bzw. 40 jährige Monatsmittel der absoluten Topographien der 1,000-mb, 850-mb- und 300-mb-Flächen sowie der relativen Topographien 500/1,000 mb und 300/500 mb über der Nordhemisphäre und ihre monatlichen Änderungen. Heft 1; Pp. 325; 103 Figs.; 18 DM. Heft 2; Pp. 168; 98 Charts; Tables; 13 DM. Heft 3. (By H. Hennig) 40 jährige Monatsmittel der absoluten Topographien der 850 mb-. 700 mb- und 300 mb-Flächen über der Nordhemisphäre für den Zeitraum 1900-1939. Pp. 55; 39 Figs.; 6 DM.

This volume, product of the School of Meteorology in Berlin under Professor Scherhag, makes an important contribution to the climatology of the upper air. Part 1 presents mean charts of the topography of the pressure levels 1,000, 850, 700, 500 and 300 mb, for a large part of the Northern Hemisphere and also the relative topographies of the layers 1,000-500 mb and 500-300 mb. The period for which these means were constructed was the five years 1949-53 and charts are given for each month and for the whole year. Monthly and annual charts of the mean temperature at 500 and 300 mb are also given. Monthly and annual means of heights of the above pressure surfaces and thicknesses of the two layers are tabulated for a large number of stations and also for the points of a grid for every 10° of longitude and 5° of latitude. The area covered by the charts is similar to that of the charts in *Die Grosswetterlagen Mitteleuropas*, that is from the pole to about 20° North over the Atlantic but extending only to about 60° North over the Pacific sector. Station tabulations are not restricted to these limits and include values for all available stations in the Northern

Hemisphere. In Part 2 average charts of the topography of the levels 1,000 and 500 mb and of the thickness of the layer 500-1,000 mb are presented for the period 1900-39 for each month of the year and for the whole of the year. The area covered is the whole Northern Hemisphere. Heights of the two levels and the relative thickness are tabulated for a grid of points. The changes in these variables from month to month are also charted. Part 3 gives similar charts of the absolute topography of the levels 850, 700 and 300 mb.

Part 1 is basic to the complete work. The average charts for the five-year period have apparently been compiled with care and using every source of information. It is regrettable, however, that the charts do not cover the whole of the Northern Hemisphere. The author has gone to considerable pains to try to eliminate errors, including systematic errors peculiar to each different type of radiosonde instrument and especially errors caused by direct absorption of radiation by the instrument. H. Heastie of the Meteorological Office, London, has recently completed a study of average upper winds over the World for the four mid-season months January, April, July and October which includes average charts of the standard isobaric levels 700, 500, 300 mb and others, for the same period 1949-53. A comparison of Fraulein Jacob's and Heastie's charts shows excellent agreement in most parts of the Northern Hemisphere except over Siberia; there Fraulein Jacob's heights are consistently lower than Heastie's, both in summer and winter; e.g., for the 300 mb level the difference is approximately 60 m in January and 40 m in July; at 500 mb the difference is about 20 m in both months. This difference is disturbing as it represents a difference in treatment of much the same data. It emphasizes the need for a determined attack on the problem of noncomparability of observations made with different types of instrument; past records must be standardized before much progress can be made in the climatology of the upper air.

The upper-air charts for the 40-yr period, in Parts 2 and 3 were constructed on the basis of the charts of average mean sea-level pressure in the well known *Normalwerte des Luftdrucks auf der Nordhemisphäre für die Periode* 1900-1939. There were, of course, no complete aerological observations for this period and the mean thicknesses of the layers were deduced from a study of relations between surface pressure and temperature and thickness in the individual months of the 5-yr period 1949-53. The construction of these upper-air mean charts for the 40-yr period must have been subjective to a certain extent but nevertheless the relationships between sea-level pressure and temperature and upper-air temperatures are sufficiently close to give considerable confidence in the final charts.

The Charts and Tables in this volume will form a standard reference for climatologists and others for many years to come.

J.K.B.

Obituary Notices

Prof. Dr. H. ARCTOWSKI

Dr. Henryk Arctowski, a Foreign Member of this Society since 1933, died on 20 February 1958 at Washington, D.C., in his 87th year. Born in Warsaw, he received his education at Belgian and French universities and after periods of specialized training in London, Paris and Zurich he was appointed to the Belgian Antarctic Expedition of 1897-99 in charge of the meteorological, oceanographical and geological work. Thus he had the distinction of being among the first men known to have endured the rigours of an antarctic winter. The vessel *Belgica* with the personnel of the expedition on board was frozen in an icefloe which for some 13 months drifted to and fro between latitudes 69° 38′ and 71° 36′S. Escape was eventually effected by means of explosives and by sawing through the solid ice night and day for several weeks. The meteorological and oceanographical observations were made for the most part by Arctowski himself, and always under his immediate supervision. His discussion of the results, published at Antwerp in 1904, was described as a model of its kind.

Dr. Arctowski's connection with the Belgian Meteorological Service ended in 1909, when he went to the United States and became head of the science division of the New York Public Library. In 1920 he was appointed to the new chair of geophysics and meteorology at the university of Lvov. In September 1939 he represented Poland at the 7th General Assembly of the International Union of Geodesy and Geophysics, in Washington. Events so turned out that he never again revisited his native country. Soon afterwards he accepted a post as research associate with the Smithsonian Institution at Washington. In that capacity he continued until his retirement in 1953. Belgium, Poland, Yugoslavia and Egypt were among the nations which awarded Dr. Arctowski honours for his contributions to geophysics and meteorology. He was at one time a recognized authority on world-wide variations of temperature, rainfall and atmospheric pressure; he also devoted much study to solar-terrestrial relationships which, he ultimately concluded after many years of scepticism, should be regarded as of paramount importance in long-term weather forecasting.

E.L.H.

Mr. I. T. NELSON

The small remaining company of this Society's Fellows who were transferred from the Scottish Meteorological Society at the fusion of the two bodies in 1921 suffered further depletion on 14 April 1958 by the death, in his 80th year, of Mr. Ian Theodore Nelson, at his home, Netherdale House, Turriff, Aberdeenshire. Mr. Nelson joined the Scottish Society in 1913. He was educated at Edinburgh Academy and Trinity College, Cambridge. After leaving the university he entered the family firm of Thomas Nelson & Sons, Publishers, Edinburgh. During the 1914-18 war he held a commission in the Cameron Highlanders and saw active service in France with the 4th Battalion. On demobilisation he returned to the publishing business, of which he became a director. Mr. Nelson resided formerly at Glen Etive, Argyllshire. For many years until about 1936 he maintained a record of the rainfall there and regularly forwarded copies of his observations to the Society.

V.N.
E.L.H.

Mr. F. TATTERSALL, B.Sc., A.Inst.P.

Mr. Frank Tattersall, of Alnwick, Northumberland, and latterly of Nairobi, who had held Fellowship in the Society since 18 May 1949, died on 4 February 1958. After a period of scholastic work for the Colonial Education Service in Jamaica he joined the R.A.F., and as a Flight Lieutenant during the 1939-45 war he undertook meteorological duties at a number of airfields in England, as well as at Inverness. He was then posted to East Africa, and from there to the Seychelles Islands. Subsequently he received a governmental appointment as scientific education officer in Kenya. Meteorology had long been among Mr. Tattersall's primary interests, and study of its manifold applications to the variety of climatic regimes found on the African continent had a particular appeal for him.

A.T.
E.L.H.

Mr. O. P. T. TURNER

Intimation has recently reached the Society that Mr. Oliver Peter Trench Turner, who had held Fellowship since 19 April 1944, died on 29 April 1957. Born at Shoeburyness in 1909, Mr. Turner received his education at Orley Farm School, Harrow, and the United Services College, Windsor. He was at one period engaged in scholastic duties and at another time worked for the British Broadcasting Corporation. But his main concerns were with agriculture and forestry. Before the 1939-45 war he inaugurated the Chiltern Tree Surgery, and latterly he had devoted his energies to a market garden at Kelso, Roxburghshire.

E.L.H.

DECCA WINDFINDING RADAR

These National Meteorological Services have ordered the DECCA W.F.1 radar:

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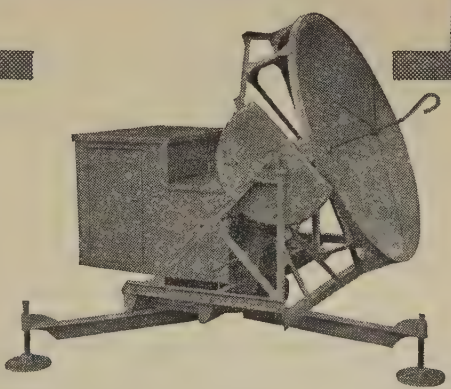
Belgian Congo	Service Météorologique
Canada	Meteorological Division, Department of Transport
Egypt	Meteorological Department
France	Office Météorologue Nationale
Lebanon	Ministère Des Travaux Publics
Mauritius	Royal Alfred Observatory
United Kingdom	Royal Society International Geophysical Year Expedition, Halley Bay, Antarctica Meteorological Office, Air Ministry
Venezuela	Venezuela Air Force—Ministry of Defence

The Decca W.F.1. is a primary radar windfinding equipment; the set is constructed specially for the measurement of winds at high altitudes and meets the need expressed by meteorologists for an accurate, cheap and robust radar which is easy to operate and requires little maintenance. A simple equipment of this type with low day to day operating costs is especially suitable for the establishment of upper wind networks. Full details of the equipment will be supplied on request.



DECCA RADAR

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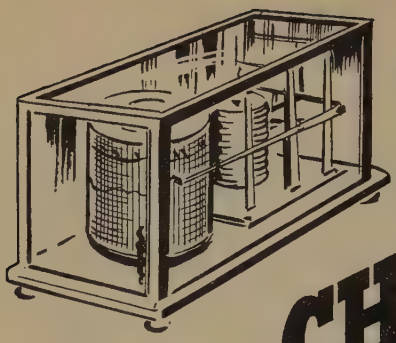

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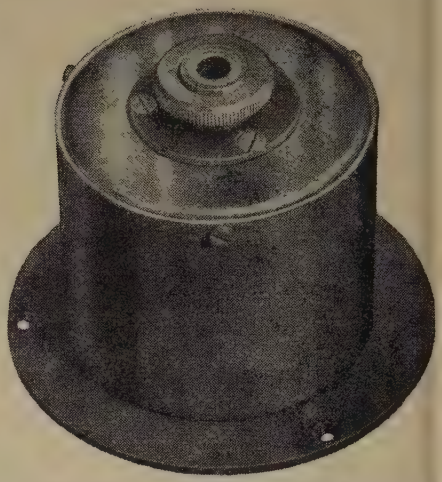

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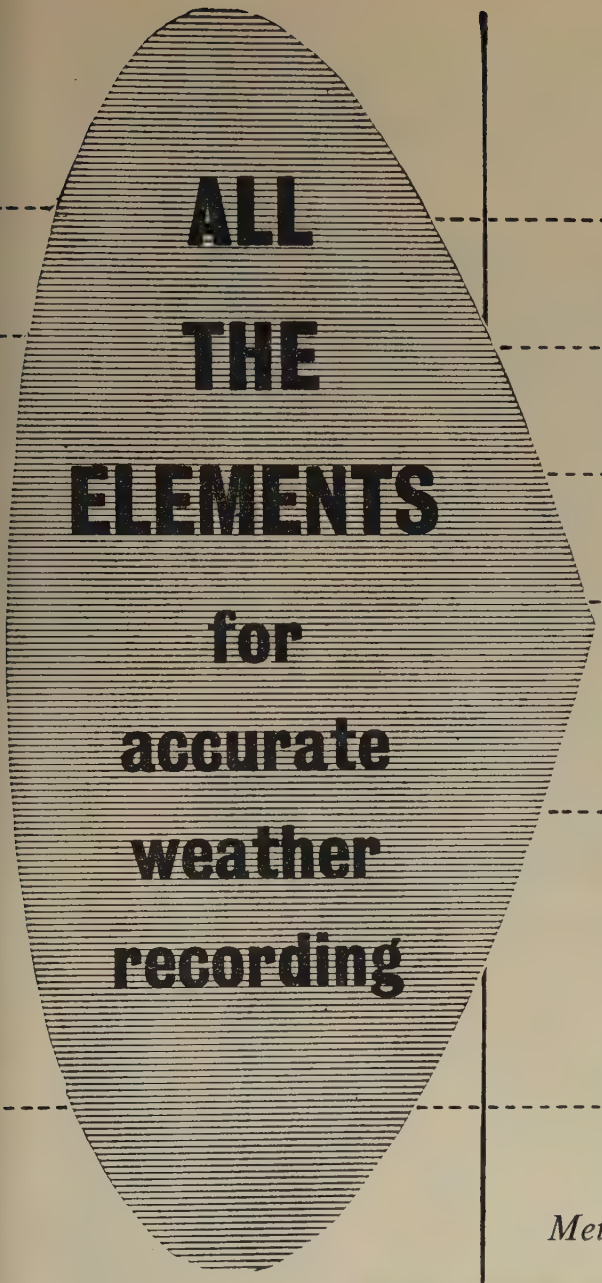
THE PHYSICAL SOCIETY EXHIBITION OF SCIENTIFIC INSTRUMENTS AND APPARATUS

Monday to Thursday, 19 to 22 January: The 43rd Physical Society Exhibition will be held at the Halls of the Royal Horticultural Society, Westminster, S.W.1.

This Exhibition is a unique one in that emphasis is placed on new developments in scientific instruments and apparatus and on the possibility of their applications in novel ways.

The opening ceremony will be performed in the New Hall by Sir Cyril Hinshelwood, F.R.S., on 19 January, at 11 a.m. Discourses will be given at 5.45 p.m. as follows: 19 January, 'The physics of controlled thermonuclear fusion' by Mr. R. S. Pease (Atomic Energy Research Establishment, Harwell); 20 January, 'Frontiers of space' by Dr. Robert Page (U.S. Naval Research Laboratory, Washington); and 21 January, 'Physical science in the Trans-antarctic Expedition' by Mr. J. G. D. Pratt, introduced by Sir Vivian Fuchs.

The Royal Meteorological Society will again be taking a small stand to display its publications. The number of the stand will be 44, and it is hoped to welcome there members and readers of the *Quarterly Journal*. Admission will be free, but by ticket only. A limited number of tickets is available on written application to the Assistant Secretary, Royal Meteorological Society.



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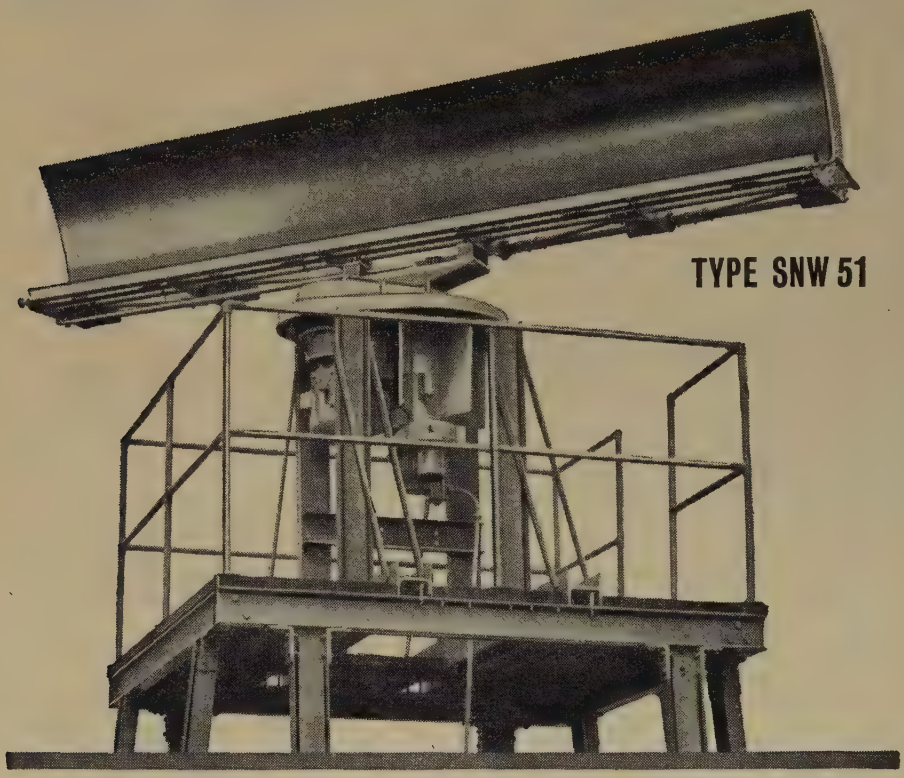
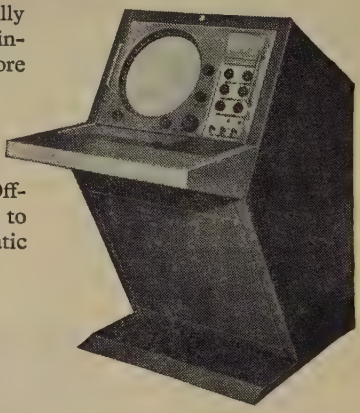


TYPE SNW 51

Marconi's Storm Detection Radar, Type SNW 51 is specially designed to present an accurate picture of storms and rain-producing clouds over ranges up to 200 n.m. It can therefore meet the needs of Meteorological Departments and Airport Authorities, particularly where it is required to derive information from inaccessible areas. Features include:— High Power Working, High-gain Tiltable Aerial (remote-controlled from display unit). Full Off-Centring Facility (enabling any sector to be expanded to cover the whole area of the tube). Remote Displays, Static or Mobile Versions available.



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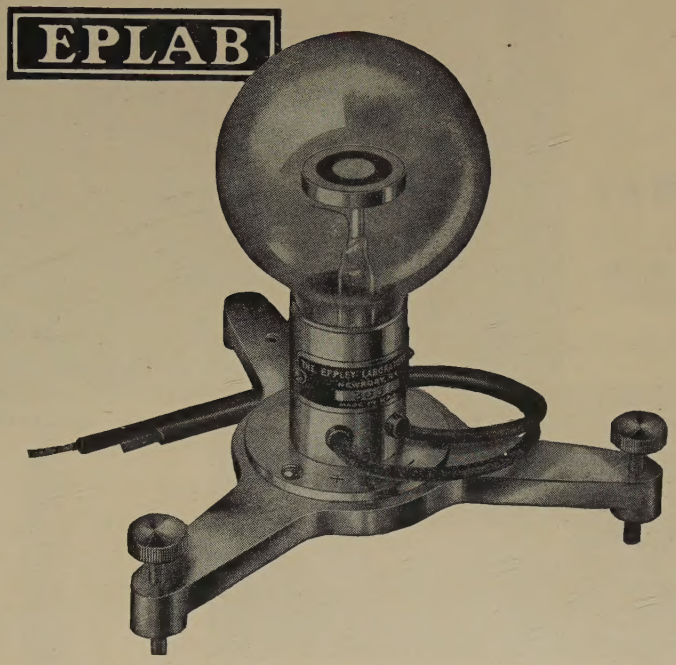

Eppley Pyrheliometers are used for solar radiation measurements at ninety-eight weather stations in the continental United States, Canada, Alaska, Greenland, Iceland, Caribbean Sea, and the Pacific Ocean. Sixty-two of these stations are under the direction of the United States Weather Bureau.

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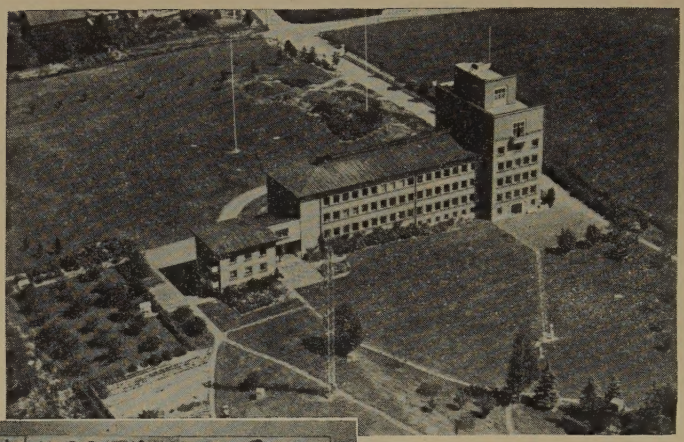

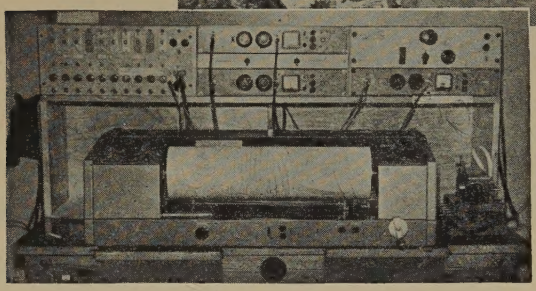

Photo: K. HARSTEAD

The Meteorologisk Institutt in Norway was founded on July 28th, 1866 and the present building at Blindern, Oslo, was erected by the Norwegian Government during the years 1936/40. The official opening was performed on November 30th, 1946, by H.M. the late King Haakon VII. The present Director of the Institute is Dr. R. Fjörtoft. The Transmitters at the Communication Centre at Blindern transmit on six different frequencies—2 long and 4 short wave—with further additional landline transmissions. To cover this schedule the Institute uses both radio and landline MUIRHEAD type D-658 Transmitters and D-649 Receivers.

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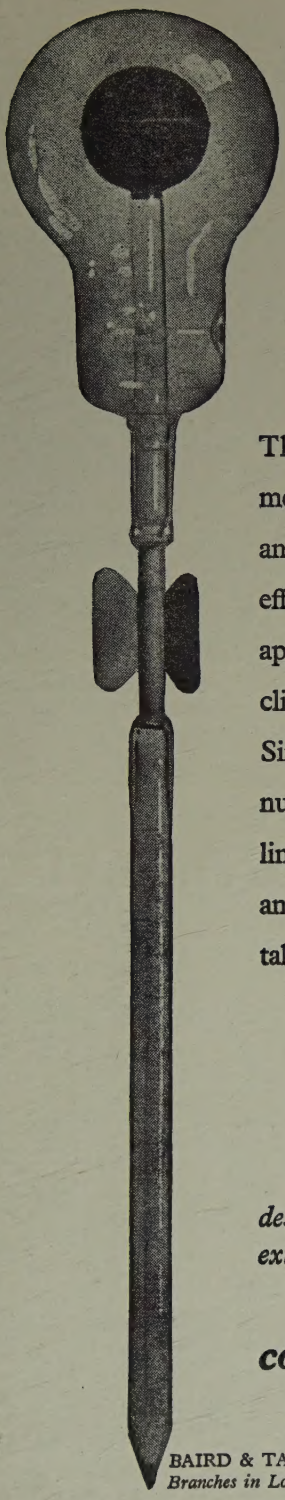

Gunn-Bellani radiation integrator

The thermal efficiency of the century-old Bellani instrument has been improved by Gunn, Kirk & Waterhouse, and later by Pereira, from 4% to over 75% thermal efficiency. In its new form, the instrument has extensive applications in meteorology, agriculture, irrigation, micro-climatology, bio-physics etc.

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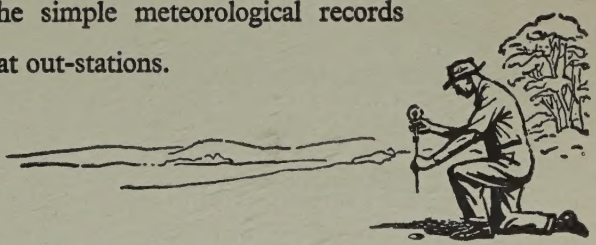

Write for Technical Leaflet No. T.86, which describes the improvements and gives the results of extensive comparisons with other instruments.

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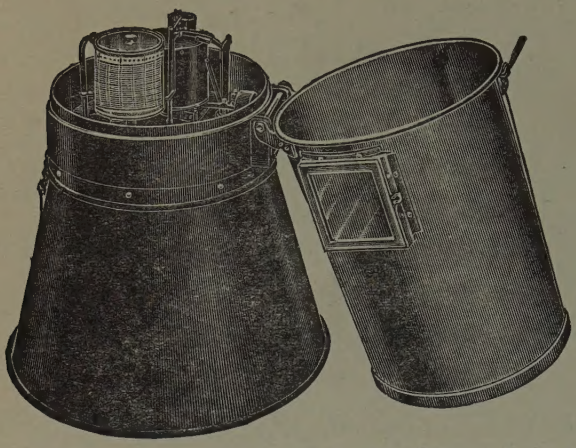

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